Volume 31 SEPTEMBER, 1961 Number 3

Journal of Sedimentary Petrology

A Publication of the Society of Economic Paleontologists and Mineralogists
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FIGS. 1–5, SEPTEMBER, 1961

SANDSTONE-PLUGGED PIPES IN THE LOWER OLD RED SANDSTONE OF SHROPSHIRE, ENGLAND[1]

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ABSTRACT

Red sandstones and siltstones of the fluviatile Ditton Series (Lower Old Red Sandstone) are pierced at many localities in the Clee Hills by sandstone-plugged pipes lying nearly normal to the bedding. These cylinders are long and narrow and sharply cut depositional structures of the host beds. The cylinders often carry siltstone fragments and sometimes course horizontally for long distances beneath siltstone layers before rising into higher beds. The less steeply inclined plugged pipes are usually associated with lateral structures formed of alternate, arcuate bands of fine siltstone conglomerate and sandstone. At one locality the cylinders rise from a convolute bedded sandstone.

The sandstone cylinders are thought to have resulted from the plugging of tubular ducts eroded by groundwater rising through unconsolidated deposits to reach the depositional surface. The lateral structures seem to indicate the migration of the loci of erosion. The Lower Old Red Sandstone plugged pipes belong to a class of structures which includes sand and mud volcanoes, spring pits, and certain cylindrical structures described from other sandstones.

INTRODUCTION

The Devonian rocks most widespread in the British Isles are of a mainly non-marine facies, long known as the Old Red Sandstone. These ostracoderm-bearing beds are well exposed in the Clee Hills of Shropshire (fig. 1) where King (1934), White (1950), Ball and Dineley (1952) and Allen (1961) have described their succession. Two divisions of the Old Red Sandstone can be recognised. The Lower Old Red Sandstone (Lower Devonian), resting sharply on marine Silurian beds, commences with the Downton Series (about 1000 ft) overlain by the Ditton Series (1200 ft). The Abdon Series (about 220 ft) followed by the Woodbank Series (about 770 ft) completes the division. Upper Old Red Sandstone (Upper Devonian) beds occur only in the south of the Clee Hills, where they lie across folded and denuded Lower Old Red rocks. Middle Old Red Sandstone strata are absent, major folding (Caledonian) and erosion taking their place.

The Old Red Sandstone succession is essentially one of red beds which reach their greatest diversity in the Ditton Series. This group is dominated by red, brown, or green sandstones and siltstones, the last the more abundant. There are small amounts of concretionary limestones and of intraformational conglomerates of siltstone and limestone debris, the rocks being known to field workers in Great Britain as varieties of cornstone (Allen, 1960). Among the sedimentary structures found in the group are cross-stratification, scour-and-fill, primary current lineation, sun cracks, rain prints, ripple marks, graded bedding, convolute bedding, flute casts, groove casts, and load casts. Large channel plugs and varied slumps are also known. All these structures promote an understanding of the sedimentation of the Ditton Series. Many are reproduced throughout the remaining Old Red Sandstone, giving the key to the deposition of these rocks.

In addition to these features, discordant sandstone-plugged pipes have been widely recognised in sandstones of the Ditton Series. These little-known structures further illuminate the Dittonian environment and are described more fully in this paper.

GENERAL DESCRIPTION

Simple Plugged Pipes

The most common plugged pipes are simple, unconnected, discordant, cylindrical structures rising more or less vertically for as much as 1 m through their host beds. Features of the lower and upper ends of the cylinders are little known. The lower ter-

[1] Manuscript received May 31, 1960.

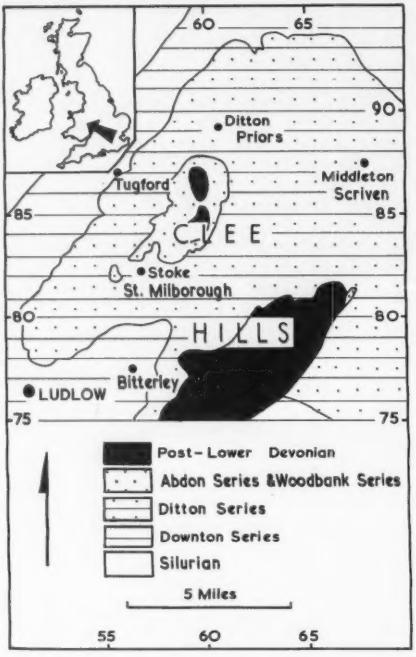

FIG. 1.—Geological map of the Clee Hills Shropshire, showing localities mentioned in the text. Coordinates of the British National Grid Reference system are shown around the border of the map. Partly based on Ball and Dineley (1952).

mination was seen at only one locality (Stoke St. Milborough), the plugged pipes originating within a convolute-bedded sandstone. The cylinders were not found beneath the contorted layer. Despite a careful search and frequent excavation, the upper ends of the sandstone plugs have nowhere been identified.

The normal hosts of the plugged pipes are clean, well-sorted, orthoquartzitic sandstones parted by thin siltstones. The sandstones are red and very fine-grained to fine-grained, normally marked either by current ripples (cuspate to parallel in plan) or horizontal lamination with primary current lineation. The red or green interleaved siltstones are often sun-cracked. Load casts and flute casts sometimes occur where sandstones sharply overlie siltstones. Less commonly the surrounding rocks include fine-grained, cross-stratified sandstones and wedging, ostracoderm-bearing intraformational conglomerates. The plugged pipes always sharply cut, but never modify, the current structures in the host beds (fig. 2).

The sandstone cylinders are usually seen on the bedding planes of the enclosing rocks (fig. 3), either as circular or elliptical bosses or shallow hollows, as erosion determines. The plugged pipes normally occur widely spaced, less commonly in scattered groups, and least often are clustered a few diameters apart. A few cylinders interlock, near Tugford for example, while sometimes one plug fully envelops another.

The least cross-sectional diameter of the plugged pipes ranges between 3 and 25 cm, the average lying near 8 cm. Though their walls are uneven in detail (fig. 2A), most of the cylinders show vertically a strikingly constant diameter. Nonetheless, some as much as double or treble their normal width to swell irregularly in the sandstone beneath a siltstone bed. A few cylinders become pinched on cutting siltstones. Most of the plugged pipes are steeply inclined or perpendicular to the bedding of the local rocks. But where interleaved sandstones and siltstones are the hosts, a few cylinders

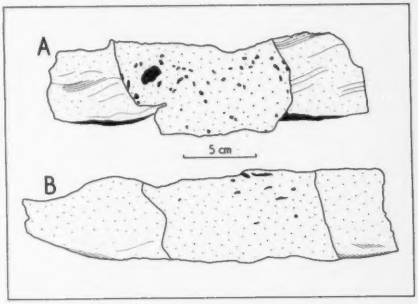

FIG. 2.—Vertical sections through simple sandstone-plugged pipes. In A an irregular cylinder of fine-grained sandstone charged with siltstone debris cuts ripple-marked, very fine-grained sandstone with small load casts at the base where it overlies siltstone; locality, Tugford 1 (Read. Univ. Dep. Geol. Sed. Res. Lab. No. S13010). In B a fine-grained sandstone cylinder pierces similarly graded sandstones with ripple marks at the finer grained base; locality, Bed 2, Tugford (Read. Univ. Dep. Geol. Sed. Res. Lab. No. S13011).

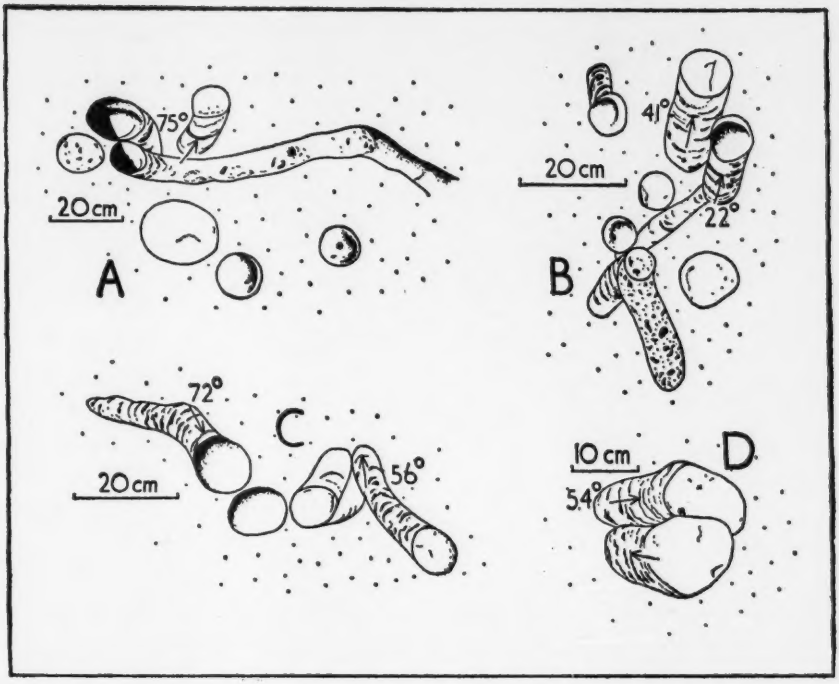

FIG. 3.—Bedding plane views of groups of simple sandstone-plugged pipes with lateral structures. With the exception of one example in Group B, all the lateral structures are of the banded variety. Drawings after photographs; the light is from the left in A, and from the upper left in B and C. All examples from Bed 2, Tugford 2.

swing away from their steep attitude in the sandstone beds to course horizontally for a short distance beneath the overlying siltstones. Beneath a siltstone at one locality (Tugford 2) (fig. 3A) a horizontal plugged pipe more than 1.2 m long branches from a steeply inclined cylinder. Eventually the plugged pipes turn from the bedding, piercing the siltstones to regain their costumary steep attitude in the next sandstone.

Sandstones, resembling the chief host beds in colour, sorting, and grade, typically form the plugged pipes. The sandstones of the plugs are normally red and very fine-grained or fine-grained, rarely being coarser than the surrounding arenites. Like the host beds, they are usually strongly cemented by calcite on earlier grain-pellicles of hematite. Locally the plugs and surrounding sandstones are lustre-mottled, the large calcite crystals extending across the boundaries of the sandstone cylinders. All the sandstone plugs abound in grains and granules of siltstone, in addition to quartz and a little exotic lithic detritus, while many are richly charged with angular siltstone pebbles (fig. 2A).

The sandstone plugs are never bedded and are normally structureless, though in some the preferred orientation of the siltstone debris is evident. In bedding plane sections the long axes of the more platy fragments often lie roughly parallel to the walls of the plugs. The same siltstone particles are steeply inclined downwards and outwards in sections along the axes of the cylinders (fig. 2A). The axially placed fragments usually lie across the plugs.

Plugged Pipes with Lateral Structures

At localities near Tugford the plugged pipes are intimately associated laterally with more complex structures. On the bedding planes these lateral structures normally appear as straight or gently curving rows of semi-circular or semi-elliptical alternate sandstone and siltstone bands, each cylinder terminating one row (fig. 3). In each structure the bands nest closely together and typically show the same curvature as the cylinder against which they fit. Thus the centres or foci are roughly colinear. The siltstone bands, often restricted to the axial portions of the structures, are usually the narrowest and the least regular. They consist of angular siltstone grains, granules, and small pebbles tightly compacted with a little quartz sand. The less common lateral structures are unbanded and homogeneous, being formed of jumbled siltstone fragments in sandstone matrices (fig. 3b). On the bedding planes both types of structure may reach as much as 65 cm in length, but an extent of between 10 and 25 cm is more usual. The lateral structures are always randomly orientated.

The banded lateral structures usually weather out as sandstone ribs and siltstone hollows inclined towards their plugged pipes at angles, ranging between 22 and 85 degrees, which are roughly constant throughout each structure. Inclinations of approximately 50 degrees were most frequently recorded at Tugford. The few sections seen normal to the general bedding confirm that in most cases the plugged pipes and the layered deposits of their lateral structures are inclined together at moderate angles. Thus the layers of sandstone and siltstone, the latter often discontinuous in depth, usually lie parallel to the walls of the cylinders. In some structures, however, opposing dips were observed, the ribs inclining away from the cylinders. In a few cases the sheets close downwards, diminishing the lateral structures in size. The layered structures often accompany plugged pipes formed of alternate steeply inclined and horizontal sections. Here the layers always parallel the steep plug and usually occupy the smaller angle made by the differently inclined cylinders.

The plugged pipes with lateral structures normally occur in groups, several of which are shown in figure 3. In A one steeply inclined cylinder turns horizontally beneath a siltstone parting, a steeply dipping lateral structure lying within the smaller angle of the bend, and cuts the lateral structure of a second plug. In group B a narrow lateral structure is invaded by two plugged pipes and the lateral structure of a third in such a way that the terminal plug is no longer evident. Groups C and D also show the transgression of one plug or layered structure by another. These relationships must indicate differences in the time of formation of the plugged pipes.

To summarise, the commonest lateral structures are formed of nested, alternate, curving sheets of siltstone detritus and sandstone. These sheets are inclined so as normally to *underlie* the associated sandstone plugs.

DESCRIPTION OF LOCALITIES

Ditton Priors

Plugged pipes were found in the Rea Brook (British National Grid Reference, SO/608904) near Derrington approximately 1 mile north of Ditton Priors (fig. 1). A large, loose block of red, fine-grained, horizontally-laminated sandstone in the stream bed carries 3 steeply inclined sandstone cylinders, each approximately 10 cm in diameter. Identical sandstones bearing traces of plugged pipes are exposed upstream. They probably lie about the middle of the Ditton Series.

Middleton Scriven

A 2 ft bed of red, fine-grained sandstone crops out in the stream (SO/672884) approximately $\frac{3}{4}$ mile northwest of Middleton Scriven (fig. 1). It is current ripple-marked and shows the molds of several plugged pipes. A few loose blocks of sandstone nearby carry 4 steeply inclined cylinders between 6 and 15 cm in diameter and up to 30 cm long.

Bitterley

In the lower Ditton Series exposed in Bensons Brook (SO/579770) approximately 1 mile south-southeast of Bitterley (fig. 1), several plugged pipes occur on the bedding planes of a red, fine-grained, horizontally-

laminated sandstone. At the stream bank follow ripple-marked, very fine-grained sandstones, siltstones with sun cracks and rubbly argillaceous limestones. The plugged pipes reach 12 cm in diameter and cut steeply up through the sandstones but were not detected in the siltstones or limestones.

Stoke St. Milborough

The Ledwyche Brook at Newton Dingle, ¾ mile southeast of Stoke St. Milborough (fig. 1), presents an excellent strike section in lower beds of the Ditton Series. At one locality (SO/574811) a red, very fine-grained, wedging, convolute-bedded sandstone reaching 9 inches in thickness is exposed for several yards. Above lie current ripple-marked very fine-grained sandstones and siltstones. Four plugged pipes, between 6 cm and 10 cm in diameter, cut these rocks and extend to the highest exposure in the stream bed. Trenching in the horizontally-laminated and ripple-marked sandstones beneath the contorted layer showed no trace of the cylinders even at the expected places of appearance.

Tugford 1

The lower Ditton Series is well exposed in the escarpment streams east of Tugford (fig. 1). Roughly 10 ft of brown and red sandstones with thin siltstones and wedging conglomerates crop out in the brook (SO /567875) approximately ¾ mile north-northeast of the village. The sandstones are very fine-grained to fine-grained and show cross stratification, scour-and-fill, current ripple mark, and primary current lineation. One siltstone is sun-cracked. The bedding planes of a sandstone about the middle of this group are extensively exposed by the stream. Over an area of approximately 10 sq mi, 23 plugged pipes were counted, some closely spaced in groups, others carrying lateral structures. Many more occur outside the area of measurement. The irregularly distributed plugs are almost all charged with siltstone pebbles. Some of the plugs have weathered to bosses but others have formed hollows. The cylinders range in diameter between 6 and 22 cm and average approximately 8.5 cm. They extend vertically or at moderate angles for 30 or 40 cm through the host beds, but their upper and lower terminations were not found. A few cylinders run horizontally for short distances, and the lateral structures reach 35 cm in length.

Tugford 2

Undoubtedly the best display of the structures is at this locality. Sandstones and thick, intraformational conglomerates carrying plugged pipes are intermittently exposed in the rivulet for approximately 800 yards downstream (south-southeastwards) of Tugford 1. Successively lower beds are descended to near the base of the Ditton Series. The most striking occurrence of the structures lies in the stream bed (SO /566871) approximately ¾ mile east of Tugford (fig. 1). A detailed plan is given in figure 4, the discordant cylinders lying within the following beds.

7. Red and green siltstones with bands and lenses of current ripple-marked, very fine-grained, calcareous sandstone, seen for approximately 2 ft.
6. Red, current ripple-marked, very fine-grained, calcareous sandstone with sun cracks near top, 6 in.
5. Red siltstone parting with widespread sun cracks.
4. Red, current ripple-marked, very fine-grained, grading up to fine-grained, calcareous sandstone, 6 in.
3. Red and green, faintly laminated siltstone, 3 in.
2. Red, horizontally-stratified, fine-grained, calcareous lenticular sandstone, siltstone locally at base, 3 to 9 in.
1. Red, horizontally-laminated and locally massive, fine-grained, calcareous sandstone with current ripple marks on top, seen for 6 in.

The cylinders reach from Bed 1 to the top of Bed 6. Excavation failed to prove the pipes in Bed 7, although their form at the top of Bed 6 suggests a further upward extension.

On the bedding planes, which total approximately 40 sq mi and are chiefly of Bed 2, 156 plugged pipes were counted. The cylinders range between 3 and 25 cm in smallest diameter and average approximately 9 cm. Almost every variety of form occurs, figure 3 showing typical representa-

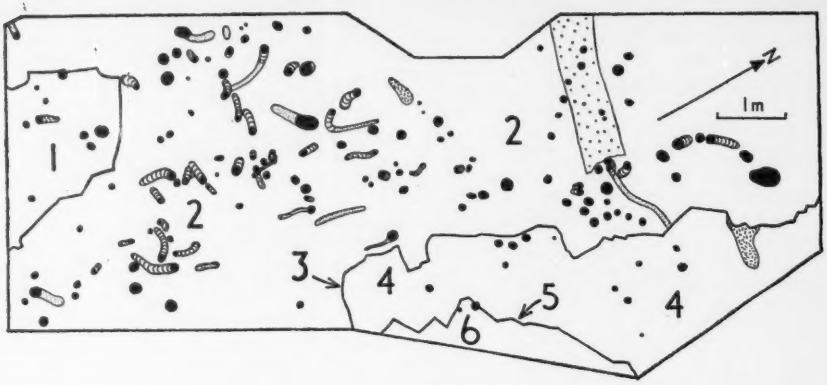

FIG. 4.—Plan of stream bed at Tugford showing distribution of sandstone-plugged pipes and lateral structures. Steeply inclined plugged pipes (black), sometimes with horizontal sections (fine stipple) just beneath siltstone layers, are widespread and many carry lateral structures, banded (shown banded) or homogeneous (stipple of short bars). The cylinders rise through interleaved sandstones and siltstones, the numbers referring to beds listed in the text. A long scour-and-fill structure (coarse stipple) crosses the stream bed and is broken by numerous pipes.

tives. Most of the plugged pipes are simple, but nearly 40 carry lateral structures, chiefly of the banded kind and rarely of the homogeneous variety. After steeply ascending the sandstones of Bed 1 and Bed 2 and before piercing higher rocks, many plugs course horizontally beneath the thick siltstone of Bed 3. Several plugs swell irregularly beneath siltstones of Bed 3 and Bed 5. The intersection of one lateral structure by another is a common feature, and the grouping of the cylinders into clusters or lines is conspicuous.

DISCUSSION

Using the many fragmentary examples studied, a general reconstruction of the sandstone-plugged pipes and their lateral structures can be presented (fig. 5). Their upper parts cannot yet be depicted. The basal relationship indicated may well be atypical, as it is based on a single record.

The plugged pipes fail to modify the form of current structures in the host beds. If the cylinders existed before the deposition of the surrounding rocks, the obstruction would almost certainly have destroyed the regular form of these current features, just as objects stranded on current-swept shoals give rise to current crescents (Peabody, 1947; Shrock, 1948, p. 130). Hence, the cylinders must postdate, or perhaps be contemporaneous with, the surrounding rocks. The plugged pipes of variable attitude certainly support the first of these suggestions, for the flat-lying plugs almost always appear beneath siltstones, showing that existing beds controlled their attitude.

Formation of the plugged pipes as deposition of the surrounding beds proceeded seems unlikely, for the bedding in the hosts does not pass into, or bend downwards towards, the cylinders. Rather, the contacts

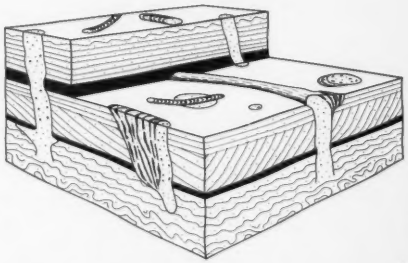

FIG. 5.—General character and relationships of sandstone-plugged pipes and their lateral structures, a reconstruction. The siltstones are shown in black.

and attitudinal changes imply the formation of the cylinders after most of the enclosing beds were laid down.

The host and plug sandstones are similarly cemented. Where the calcite cement is very coarsely crystalline, the crystals bridge the contacts between the plugs and the surrounding rocks. Clearly, the cylinders originated before the general cementation, hence while the host deposits were still incoherent.

The plugged pipes may be trace fossils representing the infilled burrows of organisms, a view supported by their simple, upright cylindrical form and frequent association with banded structures, properties typical of the burrows of many aquatic creatures (Lessertisseur, 1955). Against this explanation is the exceptionally large size of the structures and the absence of skeletal remains. Fragmentary eurypterids and ostracoderms, excepting occasional small molluscs, are alone found in the Ditton Series. Furthermore, the banded structures are in detail unlike any so far identified as organically formed.

Sand and siltstone debris, lubricated with water, might alternatively have been injected into the host sediments to form the plugged pipes. The shape and relationships of the plugs rejects this hypothesis, for branching sand veins and stringers would surely have been found had injection occurred (Shrock, 1948, p. 213–217). Instead, each cylinder appears to represent a pipe choked after a gradual and slow development to its present diameter. However, the oriented siltstone fragments of a few plugs show that occasionally the slight upward movement of lubricated detritus occurred, discounting at the same time the origin of the cylinders as down-filled, stream-scoured hollows.

The most consistent explanation of the structures supposes that tubular ducts were initiated in unconsolidated deposits by water rising to a depositional interface, and from there were gradually deepened and enlarged by the upwelling erosive flows. Choking of these ducts probably occurred whenever the velocity of the water carried through fell much below the settling velocity of any suspended detritus. Sapping and grain-by-grain erosion must have released large quantities of debris from the walls and head of each pipe during growth. Only the fines detritus could have been introduced with the water entering the bottoms of the pipes. Had the tops of the pipes been submerged, coarse debris could have been introduced into them by passing currents. Perhaps low, cratered cones surmounted the ducts to receive the detritus carried up.

The plugged pipes with horizontal sections probably record the deflection of the rising water from its steep path in the sands to a horizontal track beneath the less pervious and more cohesive siltstones. Once pierced by the water flows, the siltstones were probably eroded less readily than the loose sands above and below. Constriction of the pipes might, therefore, have resulted, the water swirling beneath the coherent layers to scour out irregular chambers. Plugged pipes at Tugford have often been observed to swell beneath such barriers or to become pinched on passing them. Erosion of the siltstones probably yielded large fragments rather than silt or clay particles. It is significant that many plugs are charged with siltstone detritus hard to distinguish petrographically from siltstones amongst the hosts.

The moderately inclined plugged pipes normally carry beneath them banded or homogeneous lateral structures. Water moving on an inclined path supposedly developed these now filled ducts. Scouring of the host deposits would have been easiest on the "hanging walls" of the pipes, causing these to migrate laterally in the direction of their downward dip. So long as a sufficient water speed was maintained, the deposition of the suspended debris on the "footwalls" would have been delayed. The filling of the resultant non-cylindrical cavity, on the final decay of water supply, might well have produced a homogeneous lateral structure, unbanded because choking took place in one operation. Such lateral structures were mapped at Tugford. But whenever the flow varied within the settling velocity range of an ill-sorted suspensate, the intermittent settling of detritus in bands of differing coarseness might be expected to occur on the "footwalls" of the pipes, a duct nonetheless re-

maining open until the final extinction of activity. Thus, both banded and homogeneous lateral structures suggest the lateral migration of pipes through incoherent beds, erosion shifting their downward facing walls while settling periodically built sideways the upward facing surfaces. Like the plugged pipes themselves, the lateral structures are scour casts.

The quantity of water needed to form each plugged pipe, to take the simpler case, may well have been large. At Tugford 2 there are 156 plugged pipes of average diameter 9 cm and of maximum length 65 cm. Assuming that 5 percent of the aqueous mixture issuing from the pipes was detritus eroded from their walls, the formation of each pipe would have required about 40 litres of water. About 6200 litres of water may therefore have passed through the pipes at this locality, equivalent to a layer about 16 cm deep over the area shown in figure 4. Flow was perhaps intermittent, leading to the full development of the pipes in a few hours or days.

The water that formed the pipes is of uncertain source. It could have been released on the compaction of beds beneath the pipes, a mechanism supported by the plugs overlying a convolute-bedded sandstone. However, this thin contorted layer could hardly have supplied the quantity of water demanded by the pipes. In no other case have the cylinders been found with slumped sandstones. More probably, the upward movement of groundwater under hydrostatic pressure created the sand-filled ducts, an ample explanation of the crowded pipes at localities like Tugford 2. Many facts of petrology, depositional structures, and general organisation combine to suggest that the Ditton Series consist in the main of flood-plain deposits. The depositional surface of such an environment is typified by a pronounced relief superimposed on a marked regional slope. Water flowing in the distributary streams, connecting with groundwater reservoirs in the contemporaneous deposits, may well have provided sufficient head for flow through the unconsolidated beds lower on the depositional surface, perhaps, as the sun cracks so commonly found with the pipes suggest, those of dried-up interdistributary basins.

OTHER RECORDS

Sandstone-plugged pipes comparable to those described above are known at other horizons and localities in the Lower Old Red Sandstone. In the Red Marls (SM/925241) and the Brownstones (SM/992201) of the Brecon Beacons, A. K. Metre and the writer discovered sandstone-filled pipes resembling in size and relationships the cylinders of the Ditton Series. The writer found the same structures in current ripple-marked green sandstones (SO/243390) at the top of the Red Marls of the Black Mountains; and a small but well exposed group of the pipes was discovered in interbedded sandstones and siltstones (SO/672185) in the Brownstones of the Forest of Dean. These examples throw no further light on the origin of the structures, but both in the Dean Forest and the Brecon Beacons the cylinders occur in a sequence marked by suncracks.

Dixon (1921, p. 29) reported from the Red Marls of Pembrokeshire, South Wales, structures "in which a sandstone fills shallow cylindrical holes eroded in an underlying marl." A photograph (A 1010) of an example from Swanlake Bay in this area is in the collection of British Photographs of the Geological Survey and Museum. The print shows sandstones and siltstones about 2 ft thick in all in a bedding plane view approximately 2.5 sq m in total extent. The beds are discordantly pierced by 30 or more closely spaced pipes filled with sandstone which range approximately between 5 cm and 20 cm in diameter and average 9 cm. The cylinders are steeply inclined relative to the bedding, and in cross-section vary from circular to elliptical according to their attitude. Several of the pipes are associated with lateral structures which extend for as much as 25 or 30 cm along the bedding planes. One of the lateral structures shown on the photograph is clearly marked by a nested internal layering very similar to that known from the pipes at Tugford 2. Hence from the above features there can be little doubt that the cylinders Dixon described are sandstone-plugged pipes comparable to the structures in the Ditton Series.

D. L. Dineley (1959, correspondence) has reported sandstone-plugged pipes from the Lower Old Red Sandstone of Freshwater West and Freshwater East, Pembrokeshire,

from the arenaceous Dartmouth Slates (Lower Devonian of continental facies) near Newton Ferrers, Devon, as well as from the Woodbay Series (Lower Old Red Sandstone) of Ekmanfjorden, Vestspitsbergen (see also Dineley, 1960, p. 21). Strikingly enough, the Spitsbergen cylinders carry mudstone flakes and occur in a facies, which also includes thick slumped sandstones, like the Lower Old Red Sandstone of the Welsh Borderland. Thus it appears that plugged pipes similar to those described in this paper are widespread in Lower Old Red Sandstone formations.

Cylindrical structures are widely known in the geological record. Gabelman (1955) described steeply inclined, sometimes interlocking or branching cylindrical structures in a siltstone from Colorado. The maroon siltstone was cross-stratified and carried intraformational conglomerates, suggesting its shallow water and perhaps continental deposition. The cylinders were attributed to the flow of water, rising or descending through the bed, just sufficient to destroy stratification in the silt without causing the transference of particles. But upward-moving, erosive currents seem a better mechanism, for Gabelman (p. 221, 222) recorded the truncation of pebbles in the host rock by the cylinder walls, the occurrence of breccias in some plugs, and the abrupt termination of the bedding in the siltstone against the cylinders. Much larger and more regular structures were reported by Hawley and Hart (1934) from cross-stratified, rippled or flat-bedded quartzites of the Potsdam Sandstone (Upper Cambrian) of Ontario. Because of mineralogical and textural similarities across the cylinders and the hosts, and the apparent grading of bedding into the featureless cores of the cylinders, the structures were explained by the action of rising water creating open packing and quicksand zones in loose sands. The cylinders lay in sandstones filling the hollows in a rugged Archean terrain. Dietrich (1952) described similar structures from the Potsdam of New York, but explained them as collapse features. Simpson (1936) noted vertical sandstone cylinders in the continental Paleocene rocks of Chubut Province, Argentina.

Genetically related to the above structures are remarkable sandstone volcanoes reported by Gill and Kuenen (1958) from marine Upper Carboniferous rocks of County Clare, Ireland. These volcanoes, normally profusely spread over the surfaces of sandstone-shale slumps or positioned on the boundary fissures of the slumped masses, were built subaqueously on bottoms swept by gentle currents as sand and silt were ejected from their vents. Sand later filled the upright feeding ducts. The water needed was thought to have been released by the settlement of the slumps. Nevill (1957) has described much smaller sandstone volcanoes from rocks of similar age in Co. Tipperary. There was here no direct association with slumped beds, and the structures were attributed to sandy water extruded during the compaction of quicksands. A solitary sandstone volcano linked with submarine faulting was recorded by Bailey and Weir (1932, p. 453) from marine Kimmeridgian beds of Sutherland, Scotland. A deep conical crater with a raised rim was reported by Thomas (Gill and Kuenen, 1958, Discussion, p. 457) from the Plateau Beds (Upper Old Red Sandstone) of the Brecon Beacons; a piedmont grading into a shallow sea was the environmental context proposed.

It is well known that during severe earthquake shocks, especially in alluvial regions, sand, water, and mud are often subaerially or subaqueously extruded along fissures or from tubular ducts. These short-lived flows build up blanket deposits or low punctate cones. Hobbs (1908, p. 121–134) and Sieberg (1908, p. 362, 363) have described and figured many examples. In many respects similar are the mud volcanoes found on the mudlumps of the present Mississippi delta. Quirke (1930) described large, shallow craters on a sand beach (river), terming them "spring pits." Coarser sand lay on their floors than on their rims. The flow of water through the sand was believed to have caused them.

Sand volcanoes and related structures have been reproduced experimentally. Rettger (1935, p. 288) formed volcanoes subaqueously in his sedimentation tank by causing the artesian flow of water through tilted beds. The low cones produced surmounted steeply inclined feeding ducts. On artificial slumps Kuenen (Gill and Kuenen, 1958, p. 454) has created miniature sand

volcanoes, observing that the cones formed as sand grains and fragments of clay settled from fountains of turbid water playing on the tops of the slumps. Thomas and Francis (Gill and Kuenen, 1958, Discussion, p. 457) report a wide variety of cones and craters produced in sand by forced jets of water. Many instances are known of sand volcanoes, often followed by caldera subsidence, formed on silty tidal flats or even sand beaches by administering mechanical shocks, such as result from jumping up and down.

CONCLUSION

The cylinders described from the Old Red Sandstone can be linked with a class of structures whose features, and modern and experimental analogues, affirm their creation by the upward movement of water through unlithified deposits. The development of these structures, broadly contemporaneous with sedimentation, seems especially to have been favoured in shallow water and continental environments, extrusion taking place either subaqueously or subaerially. On the one hand are the sand and mud volcanoes and the sandstone cylinders with fragments of the host beds, which formed when the water currents were powerful enough to erode and winnow the deposits traversed. Doubtfully on the other hand lie the cylinders with gradational contacts, and perhaps the spring pits, arising when the flows seem sufficient only to create new and less dense packings of the grains, or to winnow out the very finest particles.

The cause of the water flows is not clear, though several possibilities are likely. Field evidence sometimes indicates the release of the water from settling slumps. The compaction of quicksands has been advanced in other instances. In the many cases where there is no direct association with slumps, the tapping of groundwaters held under pressure seems a possible explanation. Seismic shocks are well able to fracture and shake down reservoir deposits and could have been an important triggering mechanism in the formation of the structures.

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The writer takes pleasure in thanking Professor P. Allen, Dr. D. L. Dineley, and Dr. R. Goldring for criticisms of the manuscript of this paper. With Dr. R. Walls, all freely gave their support, for which grateful thanks are tendered.

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FIGS. 1–4, SEPTEMBER, 1961

VARIATIONS OF SR/CA RATIOS AND MG CONTENTS IN RECENT CARBONATE SEDIMENTS OF THE NORTHERN FLORIDA KEYS AREA[1]

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ABSTRACT

The amount of strontium in Recent carbonate sediments increases from the Pleistocene reef area of the Florida Keys to the living reef area on the east. The maximum values for strontium contents are reached about one nautical mile on the leeward side of the living reef, and there is a gradual decrease closer to the living reef and seaward from it. Within the same sediments, the magnesium content decreases as the strontium content increases and reaches a maximum where the strontium content reaches a minimum. The high strontium-low magnesium content and the low strontium-high magnesium content are related to the mineralogical nature of the organic remains composing the sediments.

INTRODUCTION

The purpose of this research project was to investigate the distribution of strontium and magnesium in the Recent calcarenaceous sediments of the Florida reef area in an attempt to relate their presence and concentration to the carbonate material composing the sediments and its distance from the Pleistocene reef and the living reef.

Calculations

In the following text and in tables 1 and 2, the calcium, strontium, and magnesium contents of the sediments are reported as percent of constituents in the oven-dried, ground, and homogeneously mixed samples. The strontium to calcium atom ratios (Sr/Ca) are reported in the terms most familiar to geologists and geochemists. For example, a sediment sample from the Soldier Key Marker area contains 0.89 percent strontium and 37.4 percent calcium. The Sr/Ca atom ratio $\times 1000$ would then be:

$$\frac{0.89/87.63}{37.4/40.08} \times 1000 = 10.8$$

This indicates that for every 1000 atoms of calcium there are present about 11 atoms of strontium. By presenting the information in this way it is possible to compare the strontium content with the more familiar calcium content. The factor of 1000 is used so that the data can be represented more conveniently in larger whole units. Odum (1951) noted that the ratio changes during a process only when there is a differential action on the two elements.

Previous Work

There is a surprising lack of published material on the strontium content of Recent sediments of the type studied in this paper. Odum (1950) made a thorough analysis of the biogeochemistry of the element strontium. He reported that a Recent Bikini coral sand had a Sr/Ca atom ratio of 10.89 and that a Bikini *Halimeda* sand had a

TABLE 1.—*Strontium, magnesium, and calcium analyses of sediments of traverses T-2 to T-7 and X-1 to X-8*

Location	Ca percent	Sr percent	Mg percent	Atom Ratio Sr/Ca ×1000
T-2	36.9	0.79	0.61	9.8
T-3	37.9	0.98	0.48	11.8
T-4	36.4	1.02	0.40	12.8
T-5	37.8	1.09	0.30	13.1
T-6	37.4	0.98	0.43	12.0
T-7	37.8	0.89	0.60	10.8
X-1	34.7	0.63	0.95	8.3
X-2	36.0	0.83	0.79	10.5
X-3	35.8	0.88	0.95	11.2
X-4	36.1	0.92	0.68	11.6
X-5	37.7	1.02	0.49	12.4
X-6	38.1	1.13	0.42	13.6
X-7	37.6	0.98	0.64	11.9
X-8	37.3	0.85	0.63	10.4

[1] Manuscript received June 4, 1960.

TABLE 2.—*Strontium, magnesium, and calcium analyses of sediments exclusive of traverses*

Location	Ca percent	Sr percent	Mg percent	Atom Ratio Sr/Ca×1000
Key Largo Dry Rocks	36.5	0.86	0.51	10.8
¼ mile Northeast of Mosquito Bank Light	37.3	1.00	0.41	12.3
½ mile East of Mosquito Bank Light	37.9	0.91	0.72	11.0
Pacific Reef 180°, Caesar Creek 260°	37.6	0.91	0.49	11.1
Hens and Chickens Reef	38.0	0.96	0.47	11.5
Molasses Reef towards Mosquito Bank Light	37.5	0.88	0.59	10.7
Dixie Shoal	37.7	0.76	0.75	9.2
Conch Reef	37.4	0.82	0.69	10.0
Two miles South of Tavenier Key	37.7	1.09	0.35	13.2
Tavenier Key	36.8	0.86	0.85	10.7
Soldier Key Marker	37.4	0.89	0.44	10.9
Triumph Buoy 170°, Bache Shoal Maker 260°	36.5	0.90	0.74	11.3
½ distance from Triumph Reef to Bache Shoal	38.6	0.95	0.85	11.2

Sr/Ca atom ratio of 13.00. In 1954, Emery, Tracey and Ladd reported that the strontium content of eleven Bikini cores varied from 0.03 to 0.9 percent. Thompson and Chow (1956) found that deep sea sediments of the Indian Ocean had a Sr/Ca atom ratio of 1.94. In addition, they published that a *Globigerina* ooze from the Pacific Ocean had a ratio of 1.49 and that strontianite deposits from Washington had a ratio of 6,750.

Odum (1950, 1951), Kulp, Turekian, and Boyd (1952), and Turekian (1955) reported the strontium content of older limestones. The average Sr/Ca atom ratio of 40 limestones sampled by Odum was 1.21 and the range of 25 Paleozoic limestones 0.3 to 1.4, the average being 0.9. Kulp, Turekian, and Boyd analyzed the strontium contents of 155 limestones from Precambrian to Miocene time. The average Sr/Ca atom ratio for these samples was 0.71. The range of Sr/Ca ratios of 78 Paleozoic limestones analyzed in this study was 0.11 to 3.90, the average being 0.76.

Of the physico-chemical factors involved in determining the Sr/Ca atom ratios of shells and limestones deposited in sea water, perhaps the primary factor is the Sr/Ca atom ratio of the sea water itself. (The present Sr/Ca atom ratio in sea water is reported as 8.90, Thompson and Chow, 1956.) Turekian (1955) suggested that salinity was an important factor involved in the Sr/Ca ratio and proposed that Sr/Ca ratios of ancient sediments might serve as a key to the salinity of the seas in which they were deposited. Odum (1950) and Kulp and others (1952) also had concluded that the Sr/Ca atom ratio was related to the salinity, and in 1954 Lowenstam demonstrated that the ratio was temperature dependent in organisms having the ability to vary their aragonite-calcite contents with temperature.

Although relatively little work has been done on the strontium content of Recent sediments found in reef-forming areas, excellent studies have been made on the magnesium content of such sediments. Goldman (1926) reported the magnesium content of reef sands of the Bahamas and Thorp (1935) measured the magnesium in sands collected between Florida and the Bahamas. The most complete and most recent work of this type was done in 1954 by Chave. Chave made analyses of skeletal material of the invertebrate phyla represented in the sediments and also studied the magnesium contents of the different sizes of material composing the sediments. In all three studies the magnesium content of the sediment was related to the identifiable organic constituents of the sediment.

Ginsburg (1956) made a study of the variations in grain size and constituent composition in relation to differences of environment in the Southern Florida area but did not attempt to find chemical relations between the sediments found in the different environments. Previous to the present paper no known attempt has been made to relate the magnesium content of Recent calcarenaceous sediments to their respective strontium contents.

LOCATION OF THE AREA

Recent sediments of a clastic calcareous nature were collected from the living reef and Pleistocene reef area of the Florida Keys at 27 stations extending from Fowey Rocks (lat. 25°35′N., long. 80°06′W.) southwest to near Alligator Reef (lat. 25°52′N., long. 80°36′W.) (fig. 1).

FIELD METHODS

Samples of Recent sediments were collected from the sea floor by free diving and use of the SCUBA ("Aqualung"). The samples were placed immediately in sample bags while still under water so that portions of the sediments would not be lost in a transfer to the surface. The collecting sites were se-

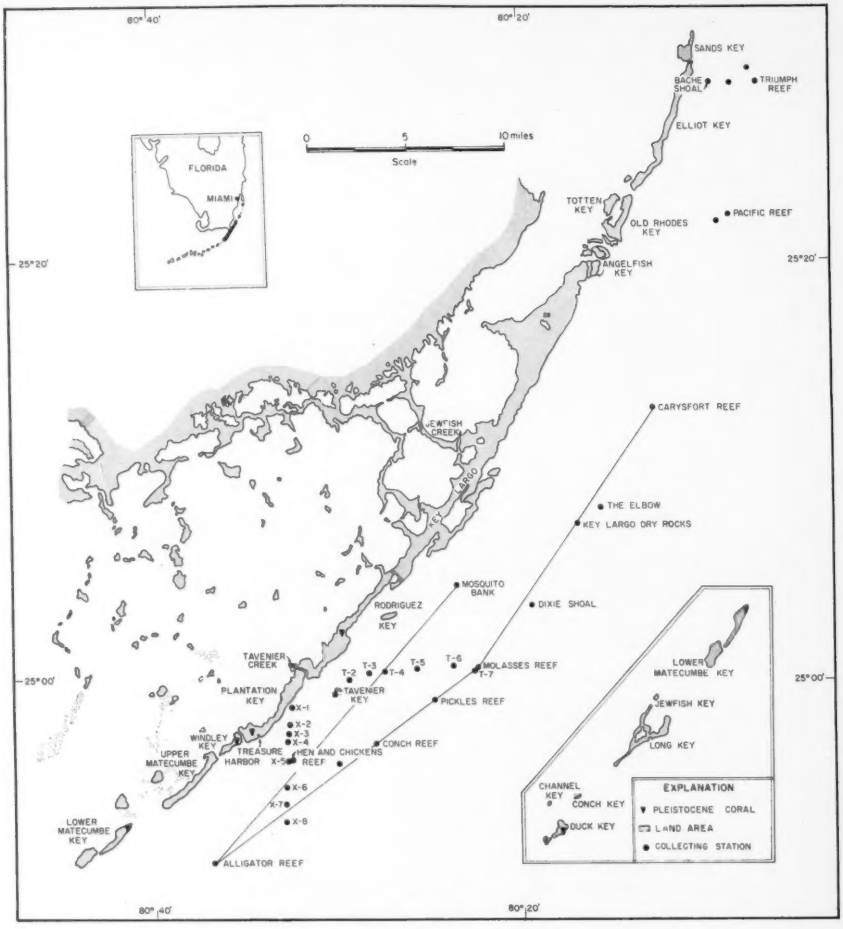

FIG. 1.—Location map of collecting stations, Florida reef area.

lected in order to be representative of the reef and back reef areas. Exact locations were determined either by dead reckoning or by using compass bearings from navigation and marker buoys.

The presence and concentration of the strontium, magnesium and calcium ions were determined by flame photometric analysis using the Zeiss Spectrophotometer PMQ II with flame equipment.

Standard Solutions

All chemicals used in this investigation were of analytical grade and met A.C.S. specifications. A stock solution of calcium, 250 parts per million (ppm), was prepared by dissolving 1250.87 mg of calcium carbonate in 10 milliliters (ml) of 6N hydrochloric acid (HCl) and diluting to 2 liters; that of strontium, 200 ppm, was prepared by dissolving 966.10 mg of strontium nitrate in 10 ml of 6N HCl and diluting to 2 liters; and that of magnesium, 200 ppm, was prepared by dissolving 1979.20 mg of magnesium sulfate in 10 ml of 6N HCl and diluting to 2 liters.

From the previously prepared stock solutions, suitable aliquots were taken and diluted for comparison standards from which calibration curves were made. To make the calibration curves, cation concentration was plotted against percent transmission. The strontium and calcium calibration curves were straight lines but the magnesium curve deviated from a straight line as more dilute concentrations were used.

Preparation of Samples for Analysis

The calcareous material making up the Recent sediments was air-dried. Gross (uncrushed) samples of the sediments were placed in porcelain crucibles, heated in an oven at a temperature of 350° C for eight hours, and cooled. At this temperature the water present was driven off and the organic matter was presumed to have been completely lost. The ash residues were powdered, homogeneously mixed, placed in sterilized vials, and stored in a desiccator. Duplicate weighed samples, 0.5 and 1.0 gm, of the homogeneously mixed ash residues were treated with 10 and 20 ml. of distilled water and 2.5 and 5.0 ml of 6N HCl respectively. These unknown solutions were quantitatively transferred to 500 ml and 1000 ml. volumetric flasks and diluted to the required volumes. The solutions were thoroughly mixed and analyzed for strontium, magnesium, and calcium, using the standard addition and dilution techniques of flame photometry after Dippel (1954).

Analysis Particulars

Strontium was determined at a wave length of 4607 angstroms using a slit width of 0.07 mm; magnesium was determined at a wave length of 3830 angstroms using a slit width of 0.07 mm; calcium was determined at a wave length of 4227 angstroms using a slit width of 0.04 mm.

Errors and Corrections

Specific substances such as organic material and other cations or anions may enhance or inhibit true transmission. To reduce these errors as much as possible, the slit widths and wave lengths were carefully selected. The interference encountered in analysis for calcium was eliminated by diluting the unknown solution to four times its volume before introducing it into the flame. Interference in the true transmission of the strontium flame was corrected by the standard addition and dilution technique. Magnesium values were corrected by using the standard addition and dilution method. However, these magnesium values were not considered as reliable as those obtained spectrochemically for the same material. A correction factor derived from a comparison of flame photometric and spectrochemical values of four samples for magnesium was applied to all samples. The corrected sample values shown in tables 1 and 2 are then comparable to the values obtained spectrochemically. Although the values of magnesium are not entirely accurate, the trends exhibited by them are considered valid. The error involved in the magnesium, strontium, and calcium values is probably of the order of 10, 5, and 5 percent respectively. In all determinations the precision of values was excellent.

RESULTS

The analyses of the sediment samples, including two traverses from the reef proper through the back reef area, revealed significant variations in both strontium and magnesium contents (tables 1 and 2). The variations are related to the distance from the Pleistocene reef which makes up the Florida Keys to the living reef 4 to 5 nautical miles away and to the material of which the sediment is composed.

The iso-strontium:calcium atom ratio map (fig. 2) illustrates that the Sr/Ca atom ratio of the sediments increase with distance offshore from the Pleistocene reef and as the living reef is approached. The Sr/Ca atom ratio reaches a maximum at a position of about one nautical mile shoreward from the main line of the living reef and decreases at the reef and seaward from it. More significant perhaps, is the magnesium content of the same sediments (Iso-magnesium percent map, fig. 3). The magnesium values decrease as the strontium contents increase and reach a minimum where the strontium contents reach a maximum. This is diagrammatically shown in figure 4.

Visual inspection of the sediments revealed that much of the finer clastic material typical of the back reef area is of an undiscernible origin. In the discernible material, the sediments found closest to the Pleistocene reef contained a great abundance of foraminiferal matter whereas samples taken successively closer to the living reef showed that the sediment is almost devoid of foraminiferal skeletons and becomes coralline in nature with much molluscan matter, some echinoid material, and a few ostracodes. The alga *Halimeda* is found in all the sediments but is most abundant where the Sr/Ca atom ratios are the highest.

CONCLUSIONS

Where the strontium content is low and the magnesium content is correspondingly high, the sediment is composed of material of a calcitic nature such as foraminiferal, echinoid, and molluscan shell remains. Conversely, the sediments in which the stron-

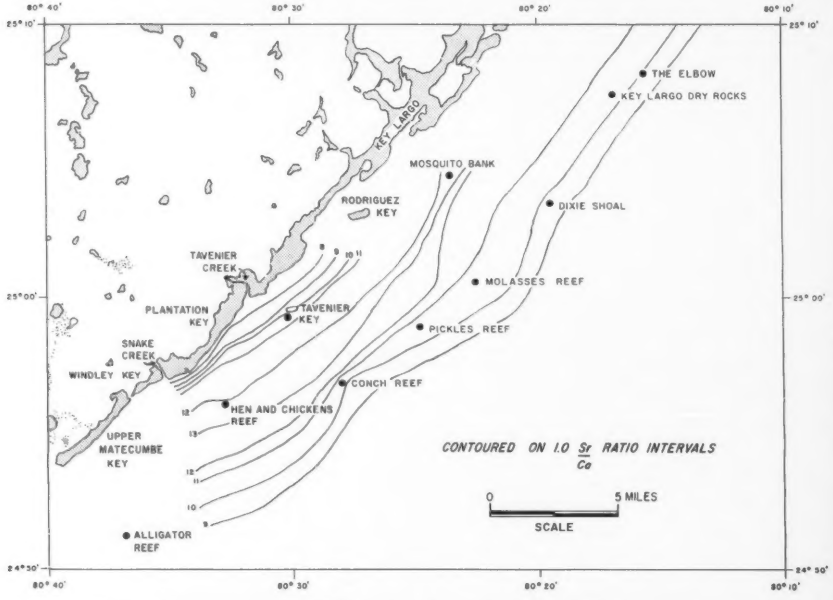

FIG. 2.—Contour map of Sr/Ca ratios of recent sediments. Florida reef area.

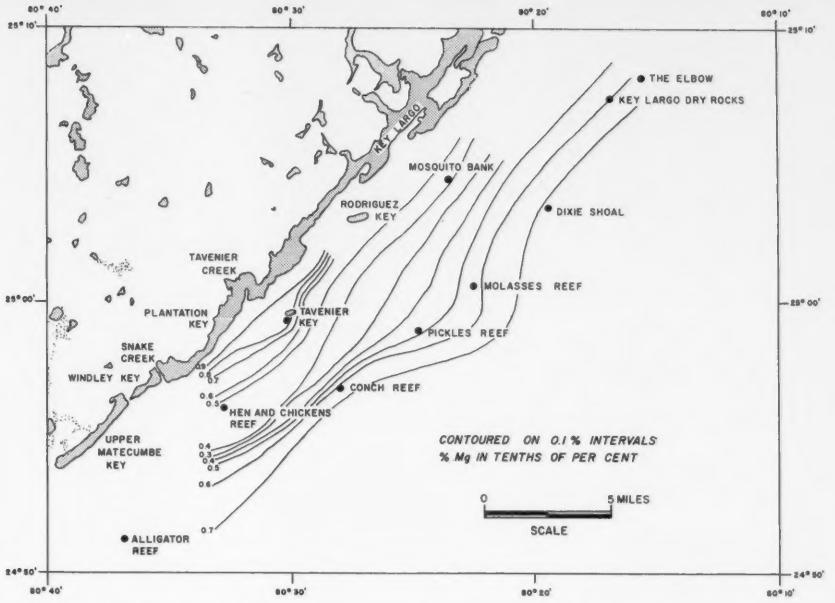

FIG. 3.—Contour map of mg percent of recent sediments, Florida reef area.

tium content is high and the magnesium content is low, is composed predominantly of aragonitic reef derived material such as coralline fragments, algae, and molluscan shells. Owing to the fact that molluscan debris may be calcite, aragonite, or mixed layers of calcite and aragonite, anomalous values may exist in some of the determinations. At locations where the magnesium and strontium contents are intermediate, the combinations of fragments composing the sediments are also intermediate. These results were not unexpected for it has been long established both in studies of laboratory precipitates and natural material (Zeller and Wray, 1956; Wray and Daniels, 1957) that aragonite will generally contain strontium to a much greater degree than will calcite. However, there has been some question as to whether magnesium prefers a calcitic crystal lattice or an aragonitic crystal lattice. The results of this study indicate that magnesium will be predominant in calcite rather than aragonite.

The previously cited data infers that the bulk of the reef derived material is deposited about one mile on the leeward side of the reef. The position of the reef derived frag-

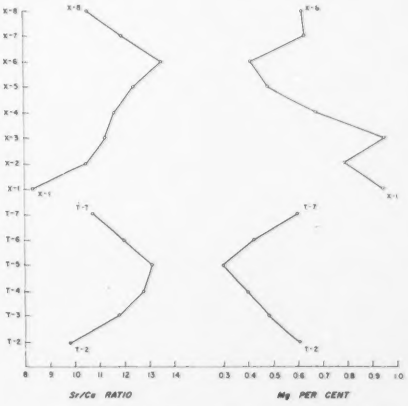

FIG. 4.—Comparison of Sr/Ca ratios and mg percent values of traverses T-2 to T-7 and X-1 to X-8.

ments behind the structure is probably controlled by the type of inter-reef currents operating in the area.

A study of the Sr/Ca atom ratios and magnesium contents of unaltered sediments, especially those with some aragonite content, might give positive information of the positions of "fossil" shorelines. In addition to this, an investigation of the strontium and magnesium contents of cored sediments in possible subsurface reef areas might prove to be a valuable exploration tool in determining the position and extent of these subsurface reefs. The strontium content of the cores might also be an indication of the position of the cores on the reef proper—that is, whether they came from the top, sides, or flanks of the reef.

In geology and geochemistry, laboratory developed ideas cannot be considered definitive until they can be applied successfully to field problems. Therefore, further research should be undertaken using cores in known subsurface reef localities where the stratigraphic relations have been fairly well established. Such areas might include the Leduc region of Alberta, Canada, and the Kirkuk region of Iraq. If the previously described strontium and magnesium relationships hold true in these areas, there would then be reason to believe that they would hold true in other subsurface reef provinces.

ACKNOWLEDGMENTS

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FIGS. 1–10, SEPTEMBER, 1961

SUBMARINE SLUMPS[1]

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ABSTRACT

Contemporary submarine slumps have previously been discussed in the geological literature primarily as a mechanism to explain otherwise anomalous topographic or sedimentary features observed on the sea-floor. In contrast, quantitative soil mechanics techniques of measuring the strength properties of sediments are used in this preliminary geological evaluation of some major marine provinces as source areas of submarine slumps and their consequent turbidity currents.

Direct shear and vane shear tests were made on undisturbed samples from the deep North Pacific, from continental and basin slopes, and from the continental shelf off California. Pre-existing strength data are summarized for comparison.

Sediments which would be theoretically stable to great thicknesses even on very steep slopes were found to predominate in all of these areas. It is concluded that the kind and rate of sedimentation required to produce unstable slopes occurs only in specialized areas of relatively rapid accumulation. These include deltaic and canyon head environments on the inner shelf, and probably also canyon or gulley walls on the open shelf and slopes. The normal open shelf and the deep-sea are believed to be nearly free of slumping.

More study of slope sediments is required particularly in and around canyons. Large scale progressive slumping offers a mechanism for failure of otherwise stable, relatively gentle slopes. The possibility of this process should be investigated in areas of suspected extensive slope failure.

INTRODUCTION

Modern concepts of turbidity currents as agents of deep-water sediment transport and deposition have resulted in the virtual abandonment of some earlier ideas of submarine slumping. Before turbidity-currents were known, for example, slumping was utilized as an agent for the transportation of displaced benthic faunas and abnormally coarse sediments; these phenomena are now generally attributed to transportation by turbid flow. Rather than decreasing the importance of slumping as a submarine processes, however, the far reaching effects of deep-water sediment transport by turbidity currents have made studies of slumping more critical than ever before. This is so because slumping is generally regarded as the cause of, or even the initial phase of, much turbidity-current action. The frequency of slumps within a given province, therefore, must largely control the activity of the province as a source of turbidity-current deposits. Furthermore it is quite possible that some of the dynamic marine processes now attributed to turbidity currents can be more realistically explained by slumping.

[1] Manuscript received June 6, 1960.

Previous geological discussions of contemporary submarine slumping have been based largely on two general lines of evidence. The first is the presence or absence of Recent sediments on various parts of the sea-floor, which the authors believed to be anomalous and therefore attributed to removal or accumulation by slumping. Probably the best known and most often quoted work of this type is a paper by Archanguelsky (1930). In his study of a series of short cores from the Black Sea shelf and slope he found that a wide band of Pleistocene sediments was exposed on the outer shelf while Recent deposits were present inshore. On the relatively steep slope beyond the shelf break he found a confused and duplicated succession of sediments which he concluded must have slumped from the outer shelf above. Because the outer shelf in this locality has approximately a 1° slope, Archanguelsky has been widely cited as having proven the process of slumping on slopes of this magnitude. Fairbridge (1947), in a paper based largely on Archanguelsky's conclusions, further suggested that the common occurrence of coarse Pleistocene deposits on the outer shelves the world over "have to some extent been kept clear of subsequent deposition by the action of gravitational slumping." Evidence to be

presented in the present paper, however, casts considerable doubt on the validity of Archanguelsky's and Fairbridge's outer shelf slumps.

The second general line of evidence is topographic. Submarine valleys, gullies, and similar features have commonly been explained as slump scars. In some cases, such as Shepard's (1955) detailed topographic study of the delta-front valleys off the Mississippi, there has been good topographic evidence for a slump origin. The valleys described by Shephard have gradients of less than 1° and extend 2 to 4 miles down slope. Aside from this gentle slope, these valleys are similar in size and configuration to some much steeper sea-gullies and sea-valleys described by Buffington (1951) and Emery and Terry (1956) on the southern California borderland basin slopes and attributed by them to possible slump action. The California sea-gullies, however, do not generally show the axial discontinuity of the delta front valleys which is characteristic of slump topography. A different type of broad landslide scar has been reported as commonly present by Gealy (1955) from studies of the topography on the upper continental slope in the northwest Gulf of Mexico. Gealy also concluded that "most of the topographic features on the continental slope (in this region) have been formed completely or modified extensively by the failure of these sediments under the stress of gravity." In sharp conflict with these ideas are reports by Kuenen (1950, p. 247), Emery and Terry (1956), and Moore (1960) of normal, extensive deposits of cohesive (argillaceous) soft Recent sediments one meter and more thick on slopes up to 18°. Obviously, there can be no rule of thumb concerning the steepness of slope upon which any sediment will accumulate. As in most things geological, each individual case must be considered separately until sufficiently understood to allow generalization.

The geological interest in the process of submarine slumping has been primarily to explain topographic or sedimentary features observed in sea floor studies. Soils engineers, on the other hand, have conventionally sought the *cause and mechanism* of particular slumps through local studies. Engineering studies of submarine slumping have, however, been essentially confined to the littoral and sub-littoral environments where mass movements of sediment have affected man-made structures. In many such areas, systematic studies of the mass properties of the sediments have been made. Terzaghi (1956) discussed marine landslides along the coast of Holland, in some Norwegian fiords, and in a sound in British Columbia. All of these slides received considerable study to determine their causes.

In studying landslides on shore or in the littoral environment, soils engineers are usually concerned with a much more complex situation than exists in the continuously deposited, non-lithified sediments of most areas of the continental shelf and slope and the abyssal sea-floor. On land the soil has generally had a more complicated history of deposition, consolidation, erosion, and weathering. Furthermore, the very important factor of pressure variations and gradients within the contained pore water must be taken into account, and the degree of saturation of the soil above the water-table is involved. Sediments of the open sea-floor present a simpler and more straightforward problem for stability analysis. All are completely saturated, are subjected to fewer changes in pore-pressure, are in general unweathered, and in most places are more homogeneous.

A general consideration of the geological significance of submarine slumping can thus be put on a sounder basis by utilizing some of the simpler quantitative methods developed by soils engineers over the past 25 years. This involves seeking the cause of slope failure both theoretically and experimentally and requires a fundamental knowledge of the mass physical properties of marine sediments deposited in the major submarine geomorphic provinces. This knowledge can be gained only through measurements of these properties by careful field sampling and laboratory testing. The most important of the properties to measure for slope-stablity analysis is the shear strength, although density, porosity, and size distribution should also be known. Other than samples from a relatively few selected coastal areas of engineering interest, very little testing of these physical characteristics of marine sediments has been done.

The present study, therefore, is intended as a preliminary sampling and measurement of strength properties of some non-lithified continental terrace and deep-sea sediments and a summary of previously existing strength data pertinent to submarine slope stability analysis. With knowledge of the range of values of these properties within a given province, one can utilize existing theory to predict the probability of slumping on slopes of known magnitude more realistically and to evaluate mass sediment movements as a factor of topographic control.

SHEAR STRENGTH

The following discussion of shear strength theory is of necessity over-simplified. Interested readers are referred to one of the several excellent standard texts on soil mechanics. Two such texts expressly written for the student as well as the working engineer are those by Taylor (1948) and Tschebotarioff (1951).

The shear strength of a sediment is a function of its cohesion plus the product of the effective load normal to the shear plane and the tangent of the effective internal friction angle. Cohesion is that portion of the shear strength which results from intermolecular and intergranular electrochemical bonding and is a characteristic property of the finer silts and clays. The angle of internal friction is a measure of intergranular rolling and sliding friction and the interlocking of grains.

In 1932 Terzaghi published results of tests on saturated clays which established the importance of pore water drainage conditions during shear. Drainage of pore water before and during shear results in quite different strength characteristics than does an undrained condition where an imposed load generates excess pore pressure. These relationships are most clearly stated in the equation:

$$\text{(1)} \qquad \text{Shear strength } \tau = C + \sigma' \tan \phi$$

where:

C = cohesion, ϕ = angle of internal friction,
σ' = effective pressure = $(N-u)$, N = normal load, and
u = pore water pressure.

Thus in a true drained condition $u = 0$, and $\sigma' = N$, hence:

$$\text{(2)} \qquad \tau = C + N \tan \phi$$

In a true undrained condition of shear $u = N$, and $\sigma' = 0$, hence:

$$\text{(3)} \qquad \tau = C$$

Many laboratory procedures have been devised to measure shear strength of soil or sediment samples under conditions of both full and zero pore water drainage. The first part of the shear strength testing program for the present study utilized the direct shear technique (fig. 1). The tests were carried out with a portable direct shear machine having a light weight plastic shear box of circular cross section, into which segments of specially selected 2 in gravity cores were placed. The upper and lower parts of the box were slightly separated, a normal load placed on the upper shear box, and a measured and increasing amount of shear stress at once applied until failure occurred along the plane of shear. This procedure was repeated at least three times with the normal stress increased each time. The maximum shear stress in each case was then plotted to define the effective strength envelope of the sediment under the conditions of the test. This type of test of cohesive sediments, in which the shear stress is applied directly after the normal load is applied, is called the "quick" test and is intended to cause an undrained condition. Another factor affecting the strength characteristics of some sediments is their capacity to expand in bulk volume during deformation. This expansion is the result of rearrangement of grain packing and is most common in densely packed sands and silts. The capacity to expand dur-

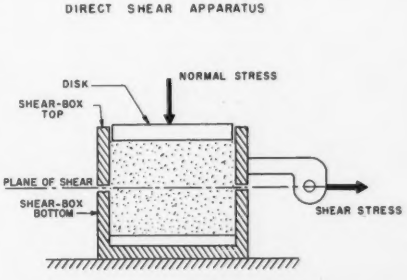

FIG. 1.—Schematic diagram of direct shear apparatus.

ing deformation effectively lowers pore pressure and may result in an undrained test giving data more representative of a partially drained condition. In this case equation (1) is more applicable than is equation (3). As will be shown later, the results of the "quick" direct shear tests of this study, in most cases, show significant angles of internal friction which would indicate a partial loss of pore pressure during shear.

Sensitivity of a sediment is defined as the ratio of the undisturbed to the remolded shear strength. This property is quite variable in natural deposits, ranging between 2 and 4 in most fine sediments and between 4 and 8 in sensitive materials. Extra sensitive and quick clays have much higher values. Sensitivity can be measured in the laboratory by successive tests of the same sample in the undisturbed state and after remolding at the same moisture content. The sensitivity of a sediment may be diagnostic in predicting its reaction to very small movements along slippage planes. In either "very sensitive" or "insensitive" sediments, however, the undisturbed shear strength must be overcome before even a very small slippage can occur on a shear plane.

STABILITY ANALYSIS, DIRECT SHEAR

Various systems of utilizing strength data for slope stability analysis are used by engineers. The system described by Taylor (1948, p. 429) for determining stability of infinite slopes (meaning those of great lateral extent as compared to the thickness of the slumped section) is a logical approach for utilizing direct shear data to study stabilities of deposits on submarine topographic features. This theory utilized the overburden pressure as the basic force which has components of shear and normal stress of a magnitude dependent on the slope angle.

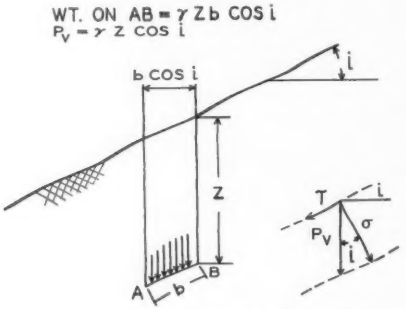

FIG. 2.—Elements of vertical pressure within a saturated sedimentary slope.

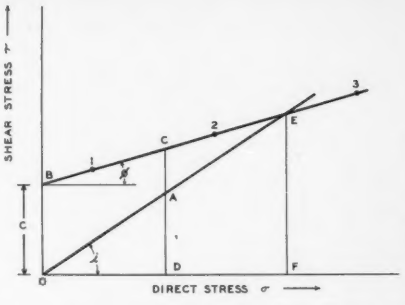

FIG. 3.—Illustrating the limited thickness to which a sedimentary layer can accumulate on a slope when the slope, i, is greater than the internal friction angle, ϕ, of the sediment.

Figure 2 is a schematic example of the forces involved on a saturated sedimentary slope of inclination, i, and thickness of sediment, Z, over a potential shear plane, AB. The weight on a strip of width b on plane AB equals the submerged density (gamma) times $Zb \cos i$, where b is the unit area. Vertical pressure, P_v, therefore equals Gamma $Z \cos i$, which is the effective weight divided by the unit area, b. As is shown in figure 2, the veritical pressure has components of shear stress (labeled tau) directed downslope, and of direct stress (labeled sigma) normal to the shear plane. The direct stress is at the angle, i, to the vertical pressure.

These same forces are shown in figure 3. The vertical pressure for a given sediment is represented here by OA; it is at the slope angle, i, to the direct stress, OD, and has the corresponding shear stress of DA. Also plotted is the strength envelope, BCE, of the given sediment. It can be seen that with a vertical pressure of less than OE the shear stress component will be less than the shear strength. For example, vertical pressure, OA, generates shear stress, DA, which is less than the shear strength, DC. The slope will

TABLE 1.—*Physical properties*

Sample	Location Lat. N.	Location Long. W.	Depth Meters	Approx. Slope Deg.	Direct Shear ϕ deg.	Direct Shear C gm/cm²	Sensitivity	Wet Dens. gm/cm³	Poros. %	Size (Wentworth) Md mm	S %	Slt %	Cl %
CK-1a	29–24		5660	3	32	14.8	5.3	1.43	73.7	.002	2	20	78
120 cm	153–06												
CK-4	42–30		5400	½	34	36.9	—	1.39	71.9	.003	4	40	56
160 cm	162–08												
CK-5	43–16		5400	3	47	19.9	—	1.30	79.3	.004	3	44	53
90 cm	163–40												
MUK-7	45–03.5		4722	0	0	19.1	3.1	1.42	74.3	.0014	0	25	75
125 cm	144–29.5												
MUK-8	45–01.4		4690	2	33	30.6	—	1.34	76.7	.0018	2	26	72
100 cm	144–28.5												
PE-1	32–41.5		2748	15	22	28.7	4.8	1.42	73.2	.0043	3	48	49
145 cm	120–18.3												
CB-5c	32–37.8		1190	0	22	17.8	4.3	1.34	78.2	.0065	0	61	39
46 cm	117–35.5												
CB-10a	32–35.4		205	5	40	15.3	4.9	1.64	62.2	.0340	21	61	18
24 cm	117–20.6												

therefore be stable. No accumulation beyond that causing a vertical pressure of *OE* is possible, so at this point the shear stress, *FE*, equals the shear strength. It is apparent from this diagram that if vertical homogeneity is assumed and the internal friction angle, ϕ, of the strength envelope is greater than the slope i then the two lines will never intersect and infinite sediment accumulation is theoretically possible.

The results of "quick" direct shear tests of deep-sea and marginal sediments are listed in table 1. It was expected that low friction angles would be found and, therefore, that only limited accumulation would be possible. However, as can be seen in table 1, seven of the eight shear tests gave friction angles (ϕ) of between 22 and 47°. These tests were made on cores from the deep North Pacific Basin, from the Patton Escarpment off Southern California which has extensive slopes of about 15 degrees, and from localities on the shelf, basin slope, and floor of the California Continental Borderland. Of particular interest are the two north Pacific samples MUK-7 and MUK-8 (fig. 4). MUK-8 is from the flanks of a low hill and MUK-7 is from the flat abyssal plain nearby. Although the density and porosity are about the same in each, the abyssal plain sample has a zero friction angle which by the infinite slope theory precludes extensive accumulation on any slope. The slope of the nearby hill, for example, is about 2° (shown by i in figure 4). The maximum thickness to which the abyssal plane sediment could accumulate on this slope is calculated as 130 cm. Hurley (1959) collected these cores and has studied the bathymetry and sediments in the locality in con-

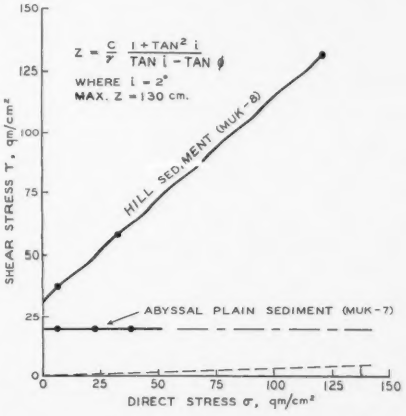

FIG. 4.—Results of direct shear tests of sediment from abyssal plain and nearby hill. Equation is shown for computing maximum thickness, Z, to which abyssal plain sediment could accumulate on adjacent hill having slope of inclination i.

siderable detail. He believes the abyssal plane sediment to be of terrigenous origin and to be relatively recent and rapidly deposited, possibly by turbidity currents. The hill sediment, on the other hand, he found to be more oxidized, to have a higher biogenous content, and to be very probably the product of slower, normal pelagic deposition.

In summary, the direct shear tests coupled with the infinite slope theory indicate the prevalence of cohesive marine sediments which would be stable to great thicknesses even on unusually steep slopes. The tests also suggest an inverse correlation between deposition rate and strength, a relationship which will be discussed in detail in a following section.

STABILITY ANALYSIS, VANE SHEAR

The analysis of slope stabilities discussed to this point has involved the use of data in which a single sample of sediment is tested for strength under varying artificially imposed loads normal to the shear plane. An alternate approach is to measure the gross shear strength within a given sediment core at various depths of burial. For this type of test in the relatively soft marine sediments in question, the most applicable device is the laboratory vane shear machine (Evans and Sherratt, 1948). An instrument of this type was constructed at The U. S. Navy Electronics Laboratory for studies of strength of loosely packed sand (Moore, 1956) and to measure the wide range of strengths in cohesive Recent marine sediments in specially collected gravity cores. The laboratory vane shear test consists of inserting a small vane into an undisturbed segment of the core and applying controlled, increasing torque until failure occurs on a cylindrical shear surface. The test is repeated at increasing depths within the core to give a consolidated-undrained shear strength gradient with burial depth or overburden pressure.

Silty clay cores from the deep North Pacific and sand-silt-clay cores collected from the San Diego Trough and slopes off San Diego were tested at various depths within the cores using the vane shear techniques. Strength data from these two areas were plotted against burial depth and least squares lines were drawn for the two sets of

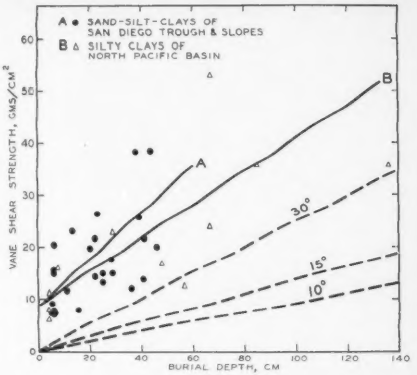

FIG. 5.—Least squares lines, A and B, of vane shear strength against burial depth from San Diego area and North Pacific basin sediments. Dashed lines show shear stress from gravity with increasing sediment thickness on slopes of 10, 15, and 30 degrees.

data (fig. 5). The correlation coefficient (r) was also computed for each of the two sets of data. The plot of the North Pacific silty clays has an r of 0.78 and the r of the San Diego Trough data equals 0.59. Using a degree of freedom equal to the number of pairs of variables less two, each of these correlation coefficients indicate a probability of less than one percent that the distribution of the plotted points could be purely random. Thus it is established that a real gradient of increasing strength with burial depth exists and that the gradients for both areas is approximately of the same magnitude.

The shear stress which would be generated by the downslope gravity component of overburden pressure on 10, 15, and 30 degree slopes was also computed and plotted as a function of burial depth along with the strength data (fig. 5). This gravity-imposed shear stress was calculated by the method described in stability analysis of direct shear data (fig. 2) and using an average figure of the bulk densities for the sediments involved. With the superposition of the strength data and calculated gravity-imposed stress data for the various hypothetical slopes it becomes apparent that the shear stress generated is well below most of the measured strengths and that as a result

of the divergence of the strength and stress lines these sediments would be theoretically stable to any thickness on slopes of up to 15 degrees and in most cases even on 30 degree slopes, assuming that the strength curves can be projected to greater depth.

Comparative data from Atlantic Ocean Floor cores, in the depth range of 400–900 fathoms, was kindly furnished by Dr. A. F. Richards of the U. S. Navy Hydrographic Office. These cores were also taken for the specific purpose of strength testing and were therefore carefully handled to maintain as nearly an undisturbed condition as possible. The strength data were obtained by the unconfined compression test. This test consists of applying gradually increasing, measured, vertical pressure to an unconfined cylindrical sediment sample and making simultaneous measurements of vertical strain. The point of maximum vertical stress equals the unconfined compressive strength. Unconfined compressive strength is generally accepted as double the shear strength value. Using this conversion, a plot of shear strength against depth was made (fig. 6) at the same relative scale as the similar plot of northeast Pacific data. The plot shows many samples of considerably greater strength than those tested for the present study. It seems a good possibility that these may be older sediments and not directly comparable to the Recent deposits tested with the vane shear machine. More significant to this discussion are the weakest sediments of the Hydrographic-Office data. These very nearly coincide with the lower limits of the abyssal north Pacific and continental borderland sediments and again indicate the probable stability of these sediments on slopes of up to nearly 30°, if the assumption of continued increase of strength with depth holds true.

It is of considerable interest in computing slope stabilities to check the validity of this assumption of the continuity of strength-depth curves at greater overburden pressures than were obtained in the relatively short gravity cores taken for the present studies. Numerous long piston cores collected in the East Pacific by the Swedish Deep Sea Expedition have been tested and described by Arrhenius (1952). Of these cores several are considered to be Pleistocene to Recent, undisturbed, and apparently continuously-deposited clays, and as such are comparable to the short cores tested for the present study. The strength tests of one 10-m core (number 51) of the Pleistocene to Recent group was chosen by Arrhenius to be representative and a least squares analysis of the strength against burial depth data was made. The strength tests reported by Arrehnius were done with the Swedish cone apparatus. The cone strength test is made by means of a metal cone, the point of which is placed immediately above the sample and is then freely dropped into the sample. By measuring the depth of penetration of the cone a "relative strength value" is obtained. This test does not give results directly comparable to the vane shear test in gm/cm² of shear strength, but through the use of con-

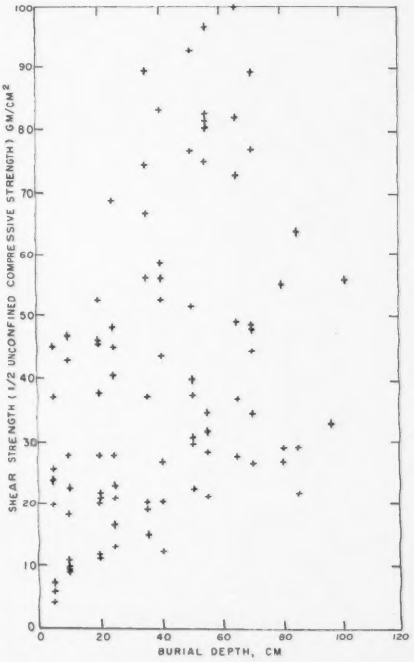

FIG. 6.—Shear strength against burial depth from Atlantic Ocean-floor cores in water depth range of 400–900 fms. Note that values of weakest sediments are comparable to those of the Pacific Ocean-floor samples.

version formulas given by Caldenius and Lundstrom (1956) the data are converted and compared to that of the present study (fig. 7).

It is interesting to note that the rate of increase in shear strength with burial depth, as computed from the Swedish Cone Test data on deep-sea sediments, is nearly identical to that derived from the vane shear methods, and that the same rate extends to the base of this impressive 10-m core. The absolute strength values of the cone test are higher, however, which may be the natural reflection of this type of test or it may be the result of the conversion factor used in changing from "relative strength" to shear strength expressed in gm/cm². Regardless of the cause of the higher strength values, the cone test data from the Swedish east Pacific cores supports the conclusion of theoretical stability of these sediments on slopes of expectable magnitude.

It is quite significant that this conclusion of predominant stability was derived from two independent systems of analysis; that is, direct-shear testing to supply data for the infinite slope theory, and measurement of vane, or equivalent, shear strength at increasing depths within the cores. The data should not be interpreted to mean that all sedimentary slopes are stable in all of the oceans, but it is strong evidence that slumps on the sea-floor in areas of normal slow deposition are at least as relatively uncommon as they are in more familiar subareal soils.

EFFECT OF RATE OF ACCUMULATION

The accumulation of sediments in the deep-sea basins is a slow process even in the vastness of geological time. In a recent summary of data from many sources Hamilton (1960) has selected 0.05 to 0.50 cm in 1000 years as being the most probable range of rates in deep-sea argillaceous deposits. Comparatively speaking, the deposits of the marginal basins and slopes have probably been rapidly formed. All of the sediments discussed thus far, however, can be considered as slowly accumulated when compared to areas of active deposition off sizable river deltas. Scruton (1956), for example, reports rates of accumulation of more than 1 foot a year, or 30,000 cm in 1000 years in the "pro-delta silty clay deposits" of the eastern Mississippi River delta.

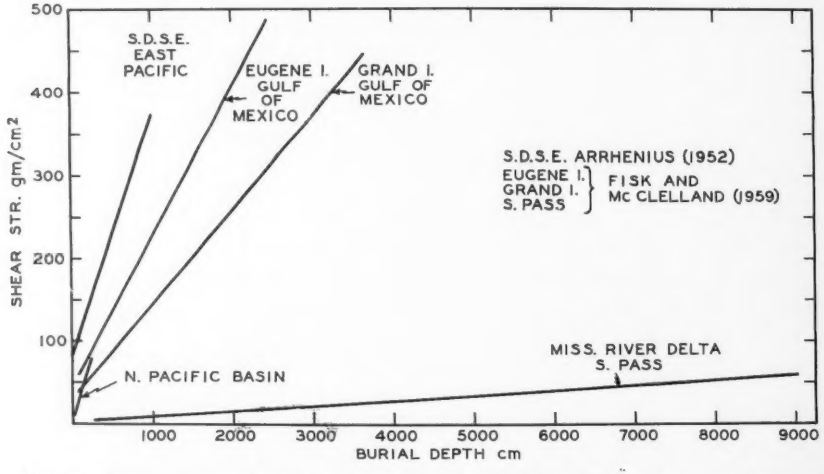

FIG. 7.—Curves of approximate shear strength against burial depth from four areas of strongly contrasting rates of accumulation. Greatest increase in strength with depth is in very slowly deposited deep-sea red clays of North and East Pacific basins; more gradual strength increases are from areas of progressively greater rates of sediment accumulation in northern Gulf of Mexico.

These tremendous variations in rate of accumulation are not without effect as a factor in determining the bulk physical properties of the sediment. Terzaghi (1956) shows on a theoretical basis that fine grained, rapidly deposited sediment may accumulate so fast that the processes of consolidation by reduction of pore water cannot keep pace. This results in excess pore water pressures within the buried sediment and does not allow the normal increase in shear strength with depth of burial. Deposits of this type are called underconsolidated clays and are normally found in the marine environment only off large rivers carrying predominately fine-grained silt and clay-sized materials. Terzaghi cited the delta-front valleys described by Shepard (1955) as probable examples of slope failure in under-consolidated sediments. Actual field testing of rapidly deposited clayey sediment off the delta of the Mississippi are described by McClelland (1956) in a discussion of foundation engineering problems connected with the construction of oil well drilling platforms. His shear-strength tests of the pro-delta deposits confirm the under-consolidated condition. More recently Fisk and McClelland (1959) have described strength tests of samples collected from a series of deep borings off the delta front and extending for approximately 150 miles to the west. Three of these borings are believed to have penetrated normal, continuously deposited sediments unaffected by subsequent erosion. These are located directly off the active South Pass of the delta, at Grand Isle approximately 50 miles west of South Pass, and at a drilling locality approximately 140 miles west of South Pass called Eugene Isle. The approximate average shear strength is plotted against depth of burial for these three borings along with the data from the deep North Pacific in figure 7. It is significant that among the three Gulf of Mexico samples there is a regular increase in strength at a given depth as the distance to the major source of sediment becomes greater. This trend is continued by the deep-sea plot. These variations in the rate of increase in strength with depth are believed to be a function of the rate of accumulation. The very low strength of the delta front deposits in comparison to the other, more slowly deposited, sediments is well illustrated, indicating the instability of this material. The instablity of the underconsolidated delta-front deposits can be more clearly shown by plotting an approximate average value of the Fisk and McClelland data for the rate of increase of shear strength with depth of burial to approximately 90 meters. This rate is then compared to the shear stress generated by the gravity component on grades of 1 and 1½ percent (fig. 8) computed by the method shown in figure 2. It is apparent from this plot that sediments of this type on a 1 percent grade are only barely stable and would not accumulate to more than about 25 m maximum on a 1½ percent grade. As these are the approximate gradients of the delta-front deposits, it is understandable that slump gullies are forming, particularly when it is considered that the actual strengths at various depths fall below the average.

Given sufficient time after a period of deposition, the excess pore pressure of an underconsolidated clay will be expended as

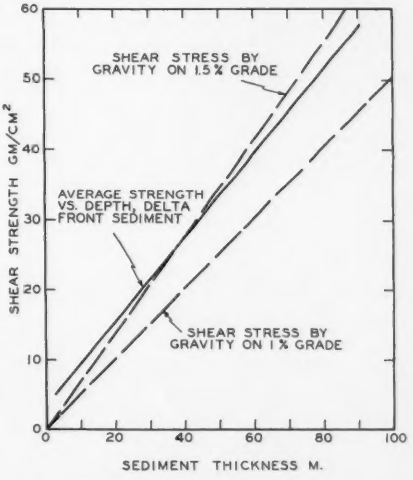

FIG. 8.—Solid line shows approximate average increase in strength with burial depth of underconsolidated delta front sediments (from data of Fisk and McClelland, 1959). Instability of these sediments is shown by dashed lines indicating shear stress from gravity on very gentle grades of 1 and 1½ percent.

water is expelled, and the deposit will become normally consolidated under its overburden pressure. The original depositonal rate, however, apparently continues to affect the mass properties of the deposit. In discussing vane shear tests of late glacial marine clayey deposits under the alluvial plane of the River Forth, near Grangemouth, England, Skempton (1948) introduced the ratio c/p. This ratio he defines as the average rate of increase in strength, c, with effective overburden pressure, p, where c equals the cohesion as determined by the consolidated undrained test which, in this instance, was achieved by the in-place vane shear technique. Skempton considered the deposit to be normally consolidated under its existing overburden pressure and noted that its ratio c/p was somewhat lower than those of several previously tested clays from other areas. He further noted that there was an apparent tendency for the c/p to decrease with decreasing values of water content, liquid limit, and clay fraction when compared to the other normally consolidated clays. There is good reason to believe that all of these sediment properties may be partially controlled by the original rate of accumulation of the deposit. Table 2 lists the ratio of cohesion, c, and overburden pressure, p, for several types of clay deposits. The Grangemouth and Koping clays are the lower and upper limits of the c/p ratios discussed by Skempton. For comparative purposes, the c/p ratios of two of the borings described by Fisk and McClelland (1959) are listed, as is the average value of the North Pacific core data of the present study. Approximate average rates of accumulation for the Gulf of Mexico sediments were calculated by assuming 10,000 years as time of accumulation and using thicknesses of the top stratum as given by Fisk and McClelland. The rate for North Pacific Basin is approximately the mean of the most probable range selected by Hamilton (1960).

TABLE 2.—*Ratio of cohesion, c, and overburden, p*

Locality	Accum. Rate cm/1000 yrs	c/p
North Pacific Basin	0.25	0.84
Koping	—	0.37
Eugene Isle	300	0.32
Grand Isle	1400	0.20
Grangemouth	—	0.17

Although the present data are insufficient to allow any firm conclusions, it is interesting to note that a semi-log plot of the ratio c/p against rate of accumulation for these three examples appears as a straight line. An extrapolation along this line would then suggest that the Grangemouth clays were deposited at a rate of about 2200 cm in 1000 years and those at Koping at less than 200 cm in 1000 years.

SUMMARY AND DISCUSSION

It has been shown that the stability of a sedimentary deposit on a given slope depends basically on the shear strength of the deposit and the rate of increase of this strength with depth of burial. These factors are controlled, in turn, by other properties such as grain size distribution, homogeneity both areally and vertically, rate of accumulation, degree of consolidation, degree of lithification, and pore-pressure conditions.

Shear strength data presented from actual measurements of sediments collected from several submarine provinces have shown unexpected stabilities particularly of the deep sea and continental and basin slope deposits. On the whole this finding is supported by strength data summarized from other sources and obtained by different testing methods. A review of published shear-strength tests for foundation studies off the Mississippi delta, on the other hand, has shown the rapidly deposited fine sediments of the delta front to be extremely unstable. Engineering literature, furthermore, contains many additional examples of slope failure in other types of coastal areas. Considering the natural variations in mass sediment properties it is not surprising that there should be strong environmental contrasts in the stability of sedimentary slopes.

The relatively small amount of strength data presently available does not allow a knowledgeable discussion of the probability of slope failure in the many and varied marine environments, but several important generalizations can be summarized from the spotty data at hand. One major conclusion is that the deep sea-floor is unlikely to produce a significant amount of slumping. The

very slow deposition of material from suspension allows the strength to build up to considerable proportions as burial proceeds, and although there are undoubtedly periods of relatively rapid deposition on some parts of the deep-sea floor by turbidity currents, these by their very nature are ponded or spread over extended areas at gradients so gentle as to preclude the possiblity of gravitational slope failure. This reasoning therefore demands that turbidity currents, as generated by slumping, are initiated in the shoaler regions of the seas; a conclusion which fits the picture of turbidity current deposits as presently described.

A general conclusion regarding the stability of the continental slope is not now possible and probably never will be because of the wide regional variation in kind and amounts of sediment and their rates of accumulation. Slope sediments vary with the width of the shelf, the nature of the source, and the complexity of the topography of the continental terrace. An area of abundant submarine canyons, or gullies, for example, should result in substantially less and therefore slower deposition on the slope, because sediments transported diagonally across the shelf and slope are trapped by the depressions where the rate of accumulation on side slopes and floors must be relatively high. The kind, amount, and rate of accumulation of sediments on the slope is also related to the extent and depth of the adjacent continental shelf. Our present eustatically "drowned" shelves are in many cases efficient sediment traps which prevent a significant amount of the detrital material introduced to the sea from reaching a resting place on the slope. Relatively large amounts of late Plesitocene to Recent sediments have, for example, accumulated on the narrow shelves off parts of southern California (Moore, 1960). During past periods when shoaler and less extensive shelves existed, there must have been a greater concentration of deposition on the slopes relative to the shelves and as a result perhaps more slumping occurred. This condition must have existed during the glacial epochs of the Pleistocene when lowered sea level exposed the shelves, and rivers dumped their loads much more directly on the upper continental slopes. The slump scars noted in the topography of the upper continental slope in the northwest Gulf of Mexico by Gealy (1955) are attributed to this rapid deposition on the upper slope and the explanation seems justified in the light of soil mechanics. Gealy's statement (quoted in the introduction of this paper) that most of the topographic features on the slope in that region are controlled by failure under gravitational stress cannot be readily accepted without measurements to back up the supposed instablity of the shales and sandstones which probably make up the bulk of the continental terrace structure in that region.

The half dozen strength measurments of slope sediments from basin slopes and the continental slope off southern California show a remarkable degree of stability. From these, it might be concluded that deposits on the basin and continental slopes of this area are generally stable. Coring in the basins has, however, shown the common occurrence of turbidity current deposits, and Gorsline and Emery (1959) attribute the origin of the turbidity currents to slumping. The explanation to this apparent anomaly probably lies in the selective activity of slumps. The slow, even accumulation of sediment on the relatively smooth surface of the majority of the slope faces results in stable deposits. Natural sediment traps in the form of gullies, canyons, and other elongate depressions trending seaward across the shelf and slope, on the other hand, accelerate sedimentation locally and must, by the same action, cause less stable or unstable deposits. The slumping which results in formation of turbidity currents and the effective distribution of the slumped material over wide areas probably recur time after time at these collecting areas while the smooth remainder of the slope remains largely quiescent.

Although submarine slumps are probably rare on the normal extensive declivities encountered on the continental slopes, it is not unlikely that local very high angle slopes which exist in areas of complex topography could accumulate sediment until they slump of their own weight on a very localized scale. These would have very little effect on the over-all regimen of slope stability unless conditions were right for the development of progressive slumping. This phenomenon of

progressive slumping has been studied in subaerial problems on several occasions and has been most satisfactorily explained by Kjellman (1955) of the Swedish Geotechnical Institute. Assume that in a slope (fig. 9) AD_1 is the sediment surface and AD is an extensive flat slip surface. The portion ACC_1 is in a state of limited equilibrium while portion CDD_1C_1 has a considerable margin of stability. A decrease in the shearing strength within part ABB_1 as a result of a rise in the pore water pressure would increase the pressure on BB_1 by some value ΔF. This would result in an equal and opposite reaction ΔF in the stable part of the slope and a compression of the portion BCC_1B_1. As a result part ABB_1 would slide down somewhat. This sliding, even though only a few inches, would cause remolding and destroy most of the shearing resistance along AB and in the adjoining parts of BC. Thus, successive failure could continue down an extensive, otherwise stable, slope. The initial rise in pore water pressure in a submarine deposit might be brought on by overriding of the upper section ABB_1 by a local slump from a high angle slope which might be triggered by an earthquake. This process of progressive slumping could cause the catastrophic rapid downslope movement of large masses of sediment and the simultaneous formation of a turbidity current which would flow well beyond the terminus of the main body of the slump.

The discussion of slumping on the continenetal shelf is more easily generalized. In the light of soil mechanics theory, existing strength measurements of open shelf sediments, and the general knowledge of the kinds of shelf sediment in many parts of the world, it appears that slumping on the normal open continental shelf is very rare. Specialized environments such as canyon heads and walls which cut into the shelf, or deltas which build out across the shelf (Shepard 1951, 1955) are necessary to produce the relatively high shear stresses and low strength required to produce slumping. Certainly a large number of known slumps have occurred on parts of the complex inner shelf and are probably more common here than in any other province of the sea, but these are essentially coastal features associated with terrigenous and neritic forces. The thinning of sediments, or absence of sediment, on the outer shelves previously attributed to slumping (Archanguelsky, 1930; Fairbridge, 1947) can be more logically explained on the basis of selective accumulation of shelf sediments. Acoustic reflection studies of sediment thickness and distribution on the shelves off California by Moore (1957, 1960) and Moore and Shumway (1959) have shown the typical sediment section to be on the order of 40 to 60 ft thick on the central shelf and to thin both shoreward and seaward from the central zone where an optimum balance between rate of accumulation and rate of deposition prevails. On wider shelves it is believed that the locus of maximum accumulation would be nearer the inner shelf. The amount of recent sediment present on the shelf varies greatly from locality to locality but in every case, on both wide and narrow shelves, it thins in a seaward direction to essentially zero either well up on the shelf or at the shelf edge depending on local supply and removal conditions. The nature of the thinning is gradual and transitional rather than abrupt, as slumping requires. Figure 10 is a reproduction of an actual acoustic reflection record from the narrow shelf off southern California at a locality approximately 50 miles southeast of Los Angeles. It shows exposed bedrock at a depth of 40 ft on the inner part of the shelf. Sediment cover is detectable in increasing amounts in a seaward direction, reaching approximately 40 ft near the central shelf and then decreasing to essentially zero near the shelf edge. This profile is typical of many obtained in this region.

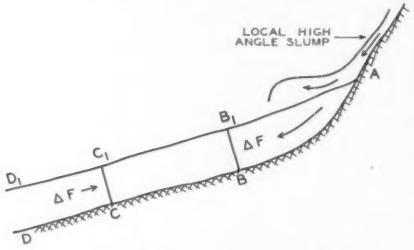

FIG. 9.—Mechanics of progressive slumping (after Kjellman, 1955).

Under certain conditions in nature, which are not yet fully understood, the deposition

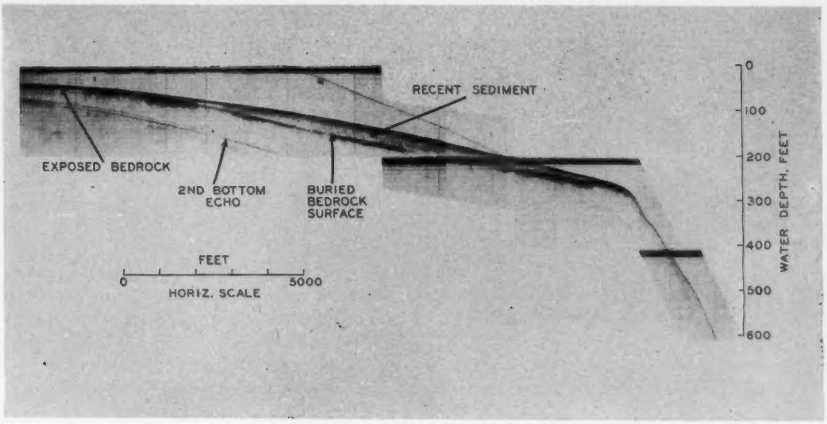

FIG. 10.—Acoustic reflection record of narrow continental shelf off southern California, showing typical distribution of Recent sediment; thinning shoreward and seaward towards shelf edge. Zones of exposed rock of Pleistocene sediments on inner or outer shelf of this type are results of differential sedimentation rather than slumping.

of fine sand- to silt-sized particles occurs in such a way as to cause the deposit to be in a metastable condition. Essentially this means that the particles are deposited in a loosely packed condition which by physical disturbance may be caused to collapse. At the instant of collapse, or liquification, the grain to grain contact is temporarily broken and the strength approaches zero. As a result of the structural collapse, the pore water pressure is temporarily increased. According to Terzaghi (1956) this excess-pore pressure may be effective in the lateral propagation of the liquification. The actual movement of a given particle within the liquified mass depends on the slope of the initial sand mass but in any case is not great, as the movement must occur during the relatively small time required for the sand to settle into a denser and more stable packing. Although Terzaghi has described slope failures attributed to collapse of metastable deposits off the coast of Holland and in Norwegian fiords, it appears unlikely that the conditions required for depositing metastable sediments are often met on the normal open shelf or slope. In general, the deposit would have to be nearly without cohesion, have its grain size in the range of 20 to 6 microns, and have a porosity of more than 50 percent. The data in table 3 are from various parts of the shelf off California in deposits believed to be of recent origin. They are otherwise randomly selected. Table 3 shows that even the surficial shelf sediments in this region have significant cohesion and are either too fine or insufficiently porous to be regarded as possible metastable deposits. A similar listing of deeper water slope materials would show even finer and more cohesive sediments. In summary, it appears that the

TABLE 3.—*Properties of surficial (0–10m) shelf sediments*

Water Depth m.	Wet Density gm/cc	Porosity (n) %	Cohesion (c) gm/cm²	Med. Diam. microns
40	1.84	49	4.0	8.8
105	1.88	46	9.2	6.7
137	1.93	42	8.7	7.2
110	1.72	62	9.1	4.7
90	1.68	60	15.2	3.3
88	1.63	65	7.0	3.3
93	1.90	46	17.7	6.2
90	1.77	52	14.8	4.7
121	1.88	51	10.6	4.1
126	1.89	47	17.8	7.7
102	2.00	69	5.8	5.8
77	1.73	55	11.0	2.2
93	1.74	58	9.2	4.4

processes of liquification of metastable sediments is probably of greatest importance in specialized, near-shore neritic or estuarine environments.

FURTHER STUDIES

Further measurements of shear strength and other physical properties of sediments are needed in and around sedimentary environments now suspected of being the locus of slumping activity. Sediments from canyon and gulley walls, for example, need to be sampled and their properties compared to those of normal intergulley or intercanyon slope sediments.

The importance of large scale progressive slumping triggered by otherwise insignificant failure of local high angle slopes needs to be determined. The mechanism of progressive slumping is theoretically sound and failures of this type have been observed on land. It may be that further soil mechanics studies in areas of suspected large scale submarine slumping will reveal this to be a major process in the modification of extensive submarine slopes.

ACKNOWLEDGMENTS

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FIGS. 1–11, SEPTEMBER, 1961

MINERALOGY AND EARLY DIAGENESIS OF CARBONATE SEDIMENTS[1]

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ABSTRACT

The mineralogy and important trace elements of Recent carbonate sediments and environmentally comparable Pleistocene carbonate rocks have been studied. All of the rocks and most of the sediments employed were obtained in southern Florida. Additional sediment samples came from the Bahama Banks, the South and East China Seas, and the Persian Gulf. Most sediments were found to consist of aragonite, high-magnesium calcite, and low-magnesium calcite. Neither dolomite nor vaterite was encountered. Shallow water carbonate sediments contain an average of about 70 percent of unstable forms of $CaCO_3$ with aragonite predominating and high-magnesium calcite dominant over low-magnesium calcite. Deep water carbonate sediments are composed predominantly of low-magnesium calcite, and high-magnesium calcite dominates aragonite. Compositional differences between shallow and deep water carbonates are believed mainly to result from differences in the importance of contributions of skeletal material by certain groups of organisms in the two environments. The mineralogy of the Pleistocene carbonate rocks studied showed them to consist mainly of low-magnesium calcite and indicated that under near-surface conditions in nature a stability sequence runs as follows: low-magnesium calcite > aragonite > high-magnesium calcite. Diagenesis of carbonate sediments is accompanied by an important loss in the level of abundance of the elements, magnesium, strontium, barium, and manganese. These elements, including magnesium, are lost from the chemical system and do not appear to form new minerals in place, although they may later do so at some other place. Since the chemical system is not closed, the significance of important volume changes which accompany diagenesis can not yet be assessed.

INTRODUCTION

This investigation was undertaken as an initial step in an effort to discover and understand some of the fundamental characteristics of carbonate sedimentary rocks. Study was concentrated on the chemistry and mineralogy of carbonate sediments and on the chemical and mineralogical changes which they undergo during alteration into carbonate rocks. The original mineralogical composition of carbonate sediments is variable and strongly influences behavior during diagenesis. Diagenetic changes, in turn, determine the nature of the resulting rocks. Knowledge of the early stages of carbonate rock formation should provide some of the basic information needed for an attack on a number of problems involving carbonate rocks.

The diagenesis of carbonate rocks, like most geological processes, requires such a long period of time that it can not be directly observed. Diagenesis may be investigated by a comparison of the carbonate sediments that are presently forming in the sea with sediments that are in various stages of alteration in the process of rock formation. This approach to the problem has been employed in the study reported here.

Carbonate sediments can apparently be derived from at least four sources: (1) the skeletal remains of marine plants and animals, (2) physicochemical precipitation from sea water, (3) bacterial precipitation, and (4) the detritus from pre-existing carbonate rocks. Only the first two sources have been found to be of general importance in normal marine environments.[4] Furthermore, in Recent carbonate sediments the bulk of the material appears to be derived from the remains of carbonate-secreting

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[4] Not considered here are limestone conglomerates which do incorporate large amounts of pre-existing rock.

organisms. Demonstrable chemical precipitates seem to be limited to ooliths, calcified fecal pellets, and the cements uniting some aggregate grains (Illing, 1954). These materials are now found to be forming in shallow water subjected to considerable diurnal temperature fluctuations.

Because the bulk of carbonate sediments is composed of the skeletal debris of organisms, the study of carbonate sedimentation must deal with the biological as well as the physicochemical influences. Investigations by Clarke and Wheeler (1922), Bøggild (1930) and others have shown that various groups of organisms have a characteristic skeletal mineralogy and employ aragonite or calcite or a combination of the two. In forms using a combination of calcite and aragonite, Lowenstam (1954) showed that the percentage of aragonite used often increases at the expense of calcite in warmer waters. Chave (1954) showed that biologically fixed calcite was of two types, one an ordinary or low-magnesium calcite, the other a high-magnesium calcite in which there is an apparently disordered substitution of more than about 4 percent $MgCO_3$ for $CaCO_3$ in the calcite lattice. He also showed that, at the taxonomic class level, the absolute amount of $MgCO_3$ present in skeletal carbonates increases with increasing water temperature.

As a result of the environmental influences exerted on the mineralogical compositon of the organically derived components of carbonate sediments, it is clear that there should be considerable mineralogical variation. Apparently this variation should be encompassed in a three-end-member system composed of aragonite, low-magnesium calcite, and high-magnesium calcite. Dolomite, which might constitute a fourth end-member, is not known to occur in significant amounts either as a chemical precipitate from normal sea-water or as a component of shells, although occasional tiny crystals (10–20μ) of supposedly authigenic dolomite have been reported by Bøggild (1912) and Correns (1937) in deep-sea sediments. A possible fifth mineral, vaterite, may occur as an ephemeral species but has not been found in significant amounts in nature.

Measurements of the proportions of the various minerals in carbonate sediments have not been generally reported in the literature, and at this time the only available information on the subject is that given by Chave (1954). An attempt has been made in the study reported here to obtain an estimate of the average mineralogical composition of carbonate sediments and of the range of variation.

In marked contrast to Recent carbonate sediments, carbonate rocks are characteristically composed only of the stable minerals calcite and dolomite or a mixture of the two. The importance of the physical and chemical conditions accompanying the conversion of mineralogically complex sediments to mineralogically simple carbonate rocks has not been fully appreciated. The conversion or loss of the minerals high-magnesium calcite and aragonite, which are both unstable under normal geological conditions, has importance in cementation and porosity problems. This same conversion of unstable species might be of significance in the formation of dolomites. An evaluation of these aspects of the diagenesis of carbonate sediments was one of the objectives of this study.

A number of elements, such as magnesium, strontium, barium, and manganese, can substitute for calcium in the $CaCO_3$ structure. Of these, magnesium is by far the most important. The diagenetic fate of the more common elements substituting for calcium in $CaCO_3$ has not been fully investigated under conditions of reasonably close geological control. As it appears possible that the concentrations of trace elements in $CaCO_3$ might influence its diagenetic behavior, the trace element content of unconsolidated sediments, partly altered sediments and carbonate rocks was investigated during the study.

LIMITATIONS OF PRESENT STUDY

The bulk of the specimens employed in this study were collected from a small geographic area, and it is therefore possible that some of the conclusions drawn may be unique to this area. As a partial check of this possibility, the mineralogy of a few samples from the South China Sea, the East China Sea, the Persian Gulf, and the Andros Lobe of the Bahama Banks was studied. These samples suggest that the results are generally valid for comparable environments

but that the mineralogy of sediments from deep and shallow environments is basically different.

It should be noted as a further qualification of the results that most of the Recent samples and all of the Pleistocene samples are believed to have been formed in shallow marine waters of more or less normal salinity and pH and in tropical to subtropical temperatures.

The diagenesis of the lithified samples studied is believed to have taken place under near-surface conditions, for geological evidence strongly indicates that these rocks were never deeply buried. It is probable that the rocks were repeatedly subjected to subaerial conditions caused by Pleistocene sea-level fluctuations.

PROCEDURE

Material Sampled

Specimens of carbonate-secreting organisms, carbonate sediments, and Pleistocene carbonate rocks were collected in southeastern Florida (fig. 10). This area is one of carbonate sedimentation today and apparently has been so during the Pleistocene and much of the Cenozoic and Mesozoic. It is characterized by vigorous reef growth along the edge of the Florida platform and by a variety of environments behind the protecting reefs and keys. Additional samples were obtained from the South China Sea, the East China Sea, and the Persian Gulf through the cooperation of Dr. K. O. Emery of the University of Southern California. Samples from the Bahama Banks (fig. 11) were obtained through the cooperation of Dr. John Imbrie of Columbia University.

Sample Preparation and Analysis

All samples studied were subjected to the same preparation procedure. The samples were ground and subjected to multiple Clorox treatments to remove included organic matter which might otherwise have interfered with the X-ray analyses and influenced trace element concentrations, and were repeatedly washed with distilled water.

Mineralogical analyses and the determinations of the percent $MgCO_3$ in high-magnesium calcite were performed by X-ray diffraction methods (see Chave, 1952). Working curves were prepared for the determination of aragonite, calcite, and dolomite. It was assumed that high-magnesium calcite has the same scattering intensity as ordinary calcite and total calcite was proportioned between the two on the basis of the relative intensity of the (100) reflection. The precision of the determinations is ± 10 percent of the amount present. The reproducibility of the $MgCO_3$ determinations in high-magnesium calcite is ± 5 percent of the amount reported. Dolomite could be detected down to a level of 1 percent of the sample.

Magnesium, barium, and manganese in the total sample were determined with an emission spectrograph. The reproducibility of these elements at the 66 percent confidence limit is: Mg ± 7.5 percent, Ba ± 12 percent, and Mn ± 14 percent. Strontium was determined by X-ray fluorescence with a precision of ± 2 percent of the amount reported. A check of the emission spectrograph determinations for magnesium by means of versene titration indicates that the determinations here reported should be considered as semiquantitative.

RESULTS

The results of all the mineralogical and chemical analyses of carbonate sediments and rocks are reported in tables 1–4.

Mineralogy of Recent Carbonate Sediments

Forty-one specimens of Recent carbonate sediments from Florida were analyzed for aragonite, dolomite, low-magnesium calcite, high-magnesium calcite, and the percent $MgCO_3$ in the high-magnesium calcite (table 1). These samples ranged from coarse reef calcarentites to very fine-grained lime muds and were collected from all of the available shallow water environments.

No dolomite was detected in any of the sediment samples. The mineralogical composition can thus be presented on a triangular diagram in terms of the remaining components, aragonite, high-magnesium calcite, and low-magnesium calcite. A plot showing the mineralogical composition of the Florida sediment samples is given in figure 1. The plots in figures 2 and 3 present the same data for sediment samples from

TABLE 1.—*Mineralogical and chemical compositional carbonate sediments from Florida*

Sample Number	% Aragonite	% Low-Mg Calcite	% High-Mg Calcite	% $MgCO_3$ in High-Mg Calcite	Dolomite	ppm Mg	ppm Sr	ppm Mn	ppm Ba	Type of Sediment	Environment
F6-1B	90	3	7	11.9	0	10,000	5600	5	13	Lime mud	Mangrove swamp
F-17-1B	61	10	20	11.8	0	17,000	4900	0	28	Lime mud	Tidal Pass
F18-1B	70	8	22	11.1	0	12,000	4300	5	12	Lime mud	Back reef lagoon
F22-16	74	3	23	15.2	0	17,000	8000	11	14	Lime sand	Reef
F22-17	80	3	17	13.3	0	11,000	6400	17	10	Lime sand	Reef
F22-18	66	2	32	15.2	0	11,000	7500	19	10	Lime sand	Reef
F22-19	71	3	26	13.0	0	9000	7600	10	14	Lime sand	Reef
F23-1B	75	0	25	14.7	0	14,000	6400	5	14	Lime sand	Reef
F24-1B	91	Tr	9	15.3	0	?	7500	5	22	Lime sand	Reef
F26-1B	88	2	10	13.0	0	9000	6200	5	10	Lime sand	Patch Reef
F26-1B	71	0	29	15.2	0	12,000	5600	5	19	Mixed seds.	Back reef lagoon
F28-1B	73	6	22	13.7	0	14,000	5400	5	11	Mixed seds.	Back reef lagoon
F29-1B	80	4	16	15.5	0	9000	6500	20	10	Lime mud	Back reef lagoon
F30-1B	65	12	23	12.8	0	18,000	4600	33	26	Lime mud	Florida Bay
F31-1B	63	15	22	12.2	0	41,000	6400	87	44	Coarse seds.	Florida Bay
F34-3B	60	16	24	12.7	0	14,000	4800	37	20	Lime mud	Florida Bay
F38-1B	42	44	14	10.5	0	9000	4000	390	11	Lime mud	Florida Bay
F40-1B	63	23	14	12.7	0	11,000	4500	40	38	Lime mud	Florida Bay
F41-1B	51	24	25	12.8	0	11,000	3700	100	17	Lime mud	Florida Bay
F42-1B	42	26	32	12.7	0	17,000	3400	70	26	Lime mud	Florida Bay
F43-1B	40	21	39	12.7	0	15,000	3700	80	12	Mixed seds.	Florida Bay
F44-1A	100	0	0	0	0	600	2000	10	11	Shell sand	Florida Bay
F45-1A	84	2	14	11.7	0	26,000	2300	45	24	Shell sand	Florida Bay
F47-1A	84	6	10	11.8	0	6000	2400	16	10	Shell sand	Florida Bay
F50-1B	42	27	31	12.7	0	17,000	3100	80	21	Mixed seds.	Florida Bay
F51-1B	63	18	19	12.7	0	19,000	4500	100	30	Lime mud	Florida Bay
F52-1B	53	22	27	13.0	0	18,000	3600	95	22	Lime mud	Florida Bay
F53-1B	56	20	24	12.4	0	14,000	3500	85	16	Lime mud	Florida Bay
F54-1B	71	13	16	12.5	0	13,000	4800	60	22	Lime mud	Florida Bay
F55-1B	57	10	33	12.1	0	14,000	2700	40	17	Lime mud	Florida Bay
F56-1B	85	5	10	12.5	0	10,000	3200	50	23	Lime mud	Florida Bay
F57-1B	90	4	6	13.7	0	2000	6100	5	18	Lime mud	Florida Bay
F58-1B	70	13	17	12.7	0	13,000	5400	50	25	Lime mud	Florida Bay
F59-1B	80	8	12	11.5	0	9000	3200	16	11	Lime mud	Florida Bay
F60-1B	76	8	16	11.8	0	11,000	6100	20	18	Lime mud	Florida Bay
F61-1B	65	12	23	13.0	0	9000	5300	35	12	Lime mud	Florida Bay
F62-1B	73	8	19	12.1	0	14,000	6000	25	21	Lime mud	Florida Bay
F63-1B	72	9	19	12.7	0	14,000	5000	30	29	Lime mud	Florida Bay
F65-1B	76	8	16	11.8	0	11,000	5200	16	14	Lime mud	Florida Bay
F73-1A	85	15	0	0	0	2000	2300	11	17	Beach sand	Outer Coast

TABLE 2.—*Mineralogical composition of some carbonate sediments from Andros Lobe of Bahama Banks*

Sample No.	% Aragonite	% Low-Mg Calcite	% High-Mg Calcite	% Dolomite	Type of Sediment	Environment
213 B	80	5	15	0	Pellet	Open Bank
317 B	76	7	17	0	Grapestone	Bank, behind Key
327 B	71	8	21	0	Grapestone	Open Bank
332 B	78	6	16	0	Pellet	Open Bank
370 B	94	1	5	0	Mud	Open Bank
376 B	85	4	11	0	Skeletal debris	Edge of Bank
392 B	99	0	1	0	Mud	Bank, behind Key
433 B	91	2	7	0	Oolite	Intertidal

TABLE 3.—*Mineralogical composition of some calcareous sediments from Persian Gulf and China Sea*

Sample No.	Area	Location	Depth	% Aragonite	% Low-Mg Calcite	% High-Mg Calcite	% Dolomite
C-4	South China Sea	17°, 26.2′ N 108°, 42′ E	96 M	0	70	30	0
C-5	South China Sea	17°, 00′ N 109°, 01′ E	112 M	0	84	16	0
C-6	South China Sea	18°, 16′ N 111°, 07.8′ E	191 M	0	77	23	0
C-7	South China Sea	18°, 58′ N 113°, 01.5′ E	448 M	0	100	0	0
C-10	East China Sea	28°, 35′ N 128°, 20′ E	1358 M	0	100	0	0
C-11	Persian Gulf	25°, 54.5′ N 55°, 07.0′ E	56 M	38	39	23	0
C-12	Persian Gulf	29°, 26.5′ N 48°, 58.5′ E	12 M	0	90	10	0
C-13	Persian Gulf	26°, 04.5′ N 52°, 11.5′ E	36 M	34	31	35	0

the Bahama Banks and from a group of sediments taken in the Persian Gulf, South China Sea, and East China Sea.

The percentage of $MgCO_3$ in the high-magnesium calcite end member of the Florida sediment samples is shown in the unique frequency curve in figure 4.

X-ray diffraction analyses of the insoluble residues from a composite sample of Florida carbonate sediments remaining after solution with Dowex 50 resin indicate that the principal impurities are quartz with small amounts of the clay minerals, illite, kaolinite, and montmorillonite. The total insolubles constitute less than 1 percent by dry weight. Microscopic investigation shows a large part of the insoluble residue to be composed of diatom and radiolarian tests and of siliceous sponge spicules.

Mineralogy of Pleistocene Carbonate Rocks

The mineralogy of 20 samples of Pleistocene carbonate rocks was determined (table 4). These rocks represent environments ap-

TABLE 4.—*Mineralogical and chemical composition of some Pleistocene carbonate rocks from Florida*

Sample No.	% Aragonite	% Low-Mg Calcite	% High-Mg Calcite	% $MgCO_3$ in High-Mg Calcite	% Dolomite	ppm Mg	ppm Sr	ppm Mn	ppm Ba	Type of Sediment
F4-1A	18	82	0	0	0	14,000	3500	5	61	Reef Debris
F14-1A	6	94	0	0	0	5000	1900	5	10	Reef Debris
F16-1A	18	82	0	0	0	3000	3000	6	10	Reef Debris
F33-1	23	77	0	0	0	4000	3300	5	10	Coquina
F36-1A	4	78	19	15.4	0	25,000	900	5	54	Reef Debris
F66-14	46	54	0	0	0	3000	5000	5	15	Coquina
F67-14	50	50	0	0	0	2000	5600	5	32	Coquina
F68-1B	56	44	0	0	0	2000	4300	5	10	Oolite and mud
F70-6B	10	90	0	0	0	5000	1800	8	10	Reef
F70-7A	73	27	0	0	0	1000	4500	5	10	Reef Sand
F71-1A	0	100	0	0	0	200	1400	5	10	Oolite and mud
F71-2A	0	100	0	0	0	500	710	5	10	Oolite and mud
F71-3A	0	100	0	0	0	500	1300	5	10	Oolite and mud
F71-4B	0	100	0	0	0	1000	1000	7	10	Oolite and mud
F77-1A	6	75	19	?	0	25,000	1200	20	29	Limy sand
F79-2B	6	94	0	0	0	600	1500	5	10	Coquina
F79-3A	33	67	0	0	0	700	1700	9	10	"Coral" sand
F82-3B	14	86	0	0	0	400	1700	5	15	Coquina
F84-1A	38	62	0	0	0	600	1400	5	16	Coquina

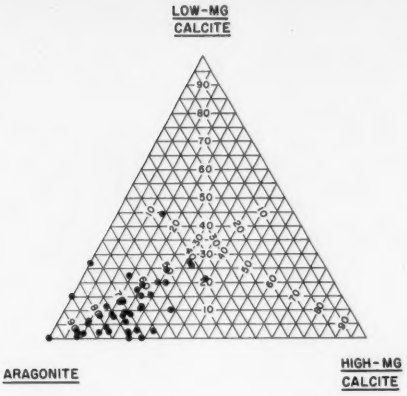

FIG. 1.—Mineralogical composition of 41 samples of Recent carbonate sediments from Florida.

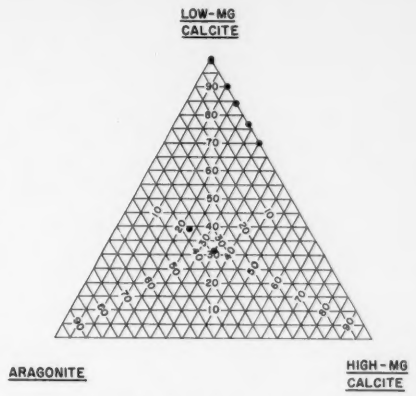

FIG. 3.—Mineralogical composition of Recent carbonate sediments from the South and East China Seas and the Persian Gulf.

parently similar to those of the modern sediments and range from reef material through coquina, calcarenite, and oolite to very fine-grained limestones. Figure 5 shows the mineralogy of these samples on a triangular diagram with the same end members used in the comparable figure for recent sediments. No dolomite was detected in any of these rocks.

Trace Element Content of Recent Sediments and Rocks

Analyses of 41 carbonate sediment samples and 20 Pleistocene carbonate rock samples from Florida were made for the abundances of the four elements, magnesium, strontium, barium, and manganese. The frequency curves in figures 6–9 indicate the distribution of these elements in the sediments and in corresponding rocks.

No trace element work was done on the samples obtained from the Andros Lobe of the Bahama Banks or on those from the

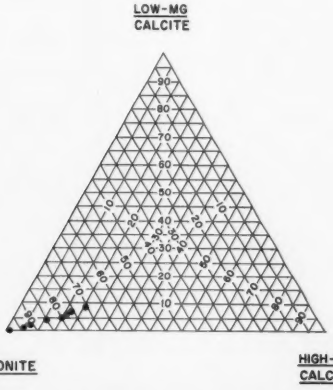

FIG. 2.—Mineralogical composition of Recent carbonate sediments from the Andros Lobe of the Bahama Banks.

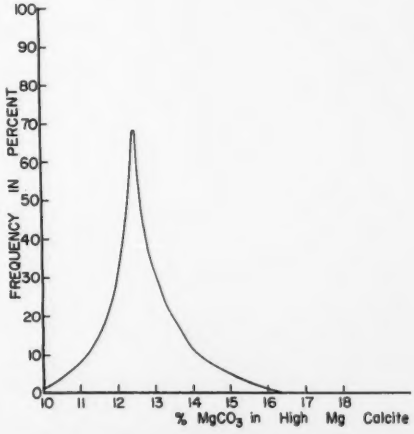

FIG. 4.—Percent of $MgCO_3$ in high-magnesium calcite in Recent carbonate sediments from Florida.

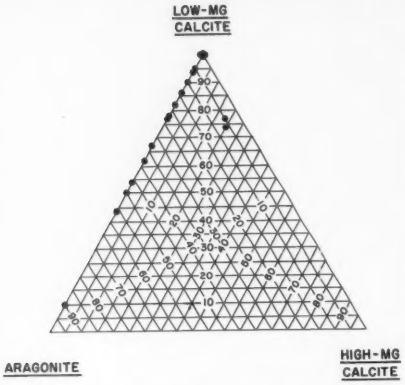

FIG. 5.—Mineralogical composition of 20 Pleistocene carbonate rocks from Florida.

Persian Gulf, East China Sea, and South China Sea because possible sources of contamination could not be evaluated and because some samples contained considerable amounts of non-carbonate sediment (tables 2 and 3).

DISCUSSION OF RESULTS

Sediment Mineralogy

Examination of figure 1 shows that aragonite is the most abundant mineralogical constituent of the sediments from Florida. It shows a tendency for high-magnesium calcite to be dominant over low-magnesium calcite. Both aragonite and high-magnesium calcite are metastable under normal geological conditions, and figure 1 shows that practically all of the sediments studied contain more than 70 percent of the metastable constituents.

The mineralogical distribution in sediments from the Andros Lobe of the Bahama Banks is similar to that found for the Florida samples. This is of significance because, unlike the sediments of the Florida area, those of the Bahama Banks contain large amounts of carbonate definitely ascribed to physicochemical precipitation by Illing (1954). Physicochemical carbonate precipitates under normal marine conditions appear to be composed of aragonite only. A sediment composed only of precipitates should therefore consist wholly of aragonite. Apparently, even where physicochemical precipitation

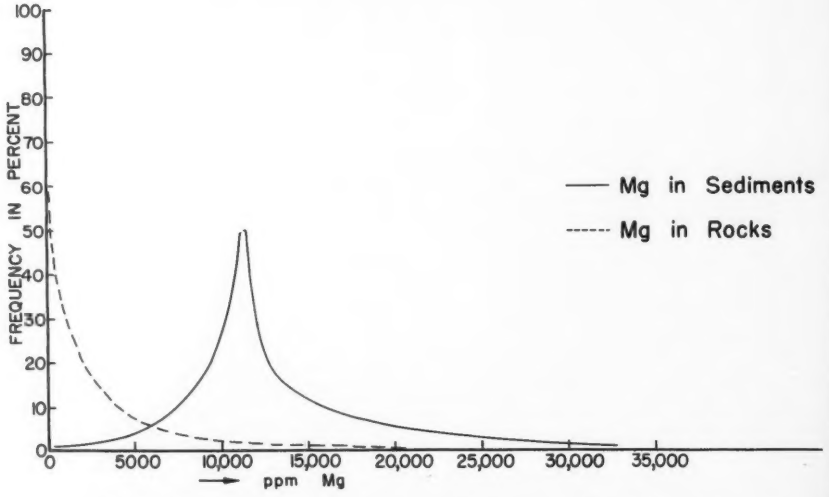

FIG. 6.—Magnesium concentration in ppm in Recent carbonate sediments and Pleistocene carbonate rocks from Florida.

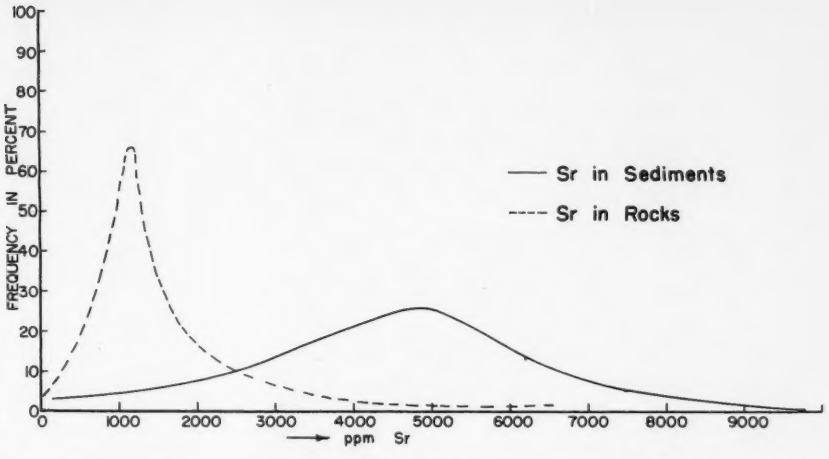

FIG. 7.—Strontium concentration in ppm in Recent carbonate sediments and Pleistocene carbonate rocks from Florida.

predominates, admixture of skeletal debris is often sufficiently common to produce a sediment composed of three end members.

Analyses of eight samples from the South and East China Seas and from the Persian Gulf are shown in figure 3. These samples come from a variety of ecological conditions different from those sampled either in Florida or in the Bahama Banks. Three samples are from the Persian Gulf and were

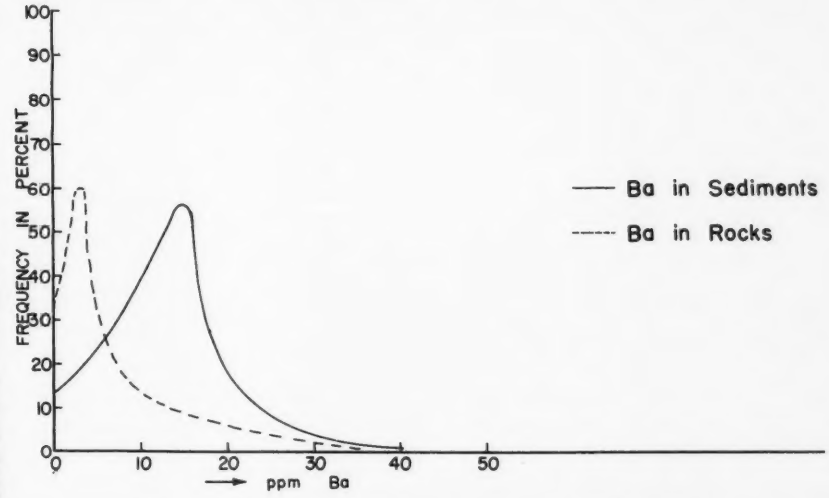

FIG. 8.—Barium concentration in ppm in Recent carbonate sediments and Pleistocene carbonate rocks from Florida.

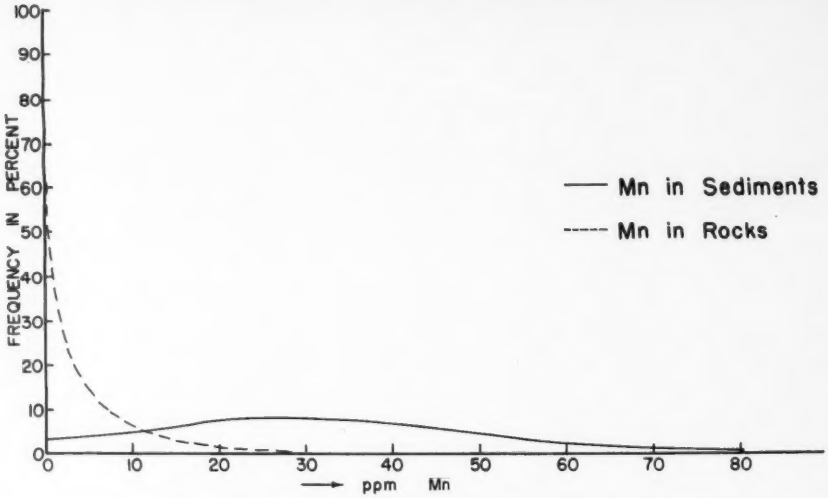

FIG. 9.—Manganese concentration in ppm in Recent carbonate sediments and Pleistocene carbonate rocks from Florida.

taken in waters less than 200 ft deep. This is a relatively shallow environment though much deeper than either the Florida or Bahama samples. Two of the samples show compositions consisting of roughly equivalent amounts of all three end members and thus fall near the center of the graph. The other sample consists primarily (90 percent) of low-magnesium calcite, indicating derivation from skeletal material. Microscopic examination shows the sample to consist largely of planktonic foraminifera which under the existing conditions apparently secrete low- rather than high-magnesium calcite.

A single sample from the East China Sea was taken at a depth of 4455 ft and consists exclusively of low-magnesium calcite. Microscopic examination shows that the carbonate fraction appears to be almost entirely made up of the tests of planktonic foraminifera.

Four samples from the South China Sea were examined. One was taken at 1469 ft and consists only of low-magnesium calcite. This mineral appears to be contributed by the tests of planktonic foraminifera. The remaining three samples were taken in less than 630 ft of water. Each consists of a mixture of low- and high-magnesium calcite with the low-magnesium form dominant. No aragonite was found in any of the China Sea samples. Microscopic examination shows them to consist primarily of foraminiferal tests so that little or no aragonite would be expected.

Comparison of all of the mineralogical analyses from this study suggests that carbonate sediments produced in warm, shallow water either by biochemical activity or a combination of biochemical and physicochemical processes show a well-defined composition in which aragonite predominates and in which high-magnesium calcite is dominant over low-magnesium calcite. These sediments are generally composed of more than 70 percent unstable carbonate minerals. The results of this study indicate that carbonate sediments indigenous to deep water are composed predominantly of low-magnesium calcite, and that high-magnesium calcite dominates over aragonite. Deep water carbonates are, therefore, more stable mineralogically than those of shallow water environments. This mineralogical difference between shallow water and deep water carbonate sediments, if real, must reflect the mineralogical habit of the groups of organisms whose skeletons contribute

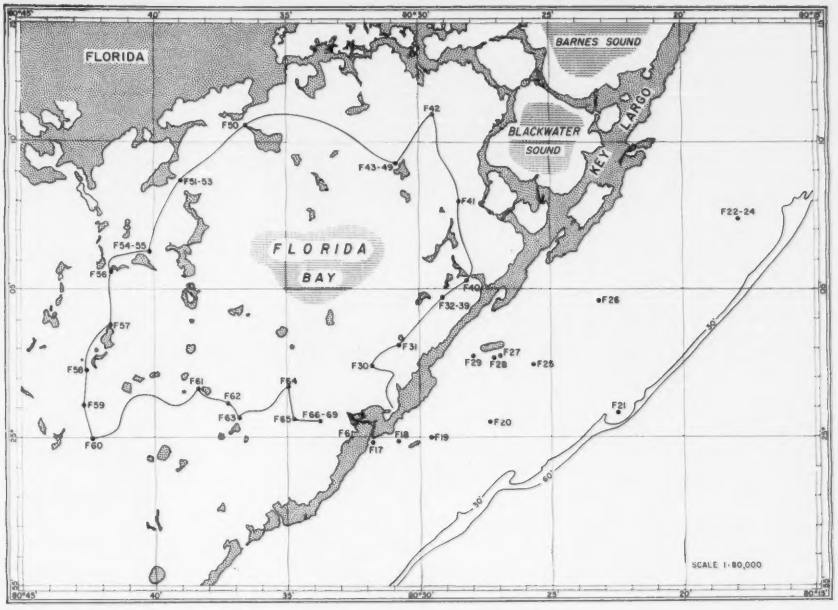

FIG. 10.—Sampling stations in Florida.

importantly to the sediment. Many exceptions to this general picture are to be expected, in part because of local concentrations of particular groups or organisms and in part because of transport of shallow water sediments into deeper water. Because shallow water carbonate deposits tend to consist largely of unstable carbonate minerals, while deeper water sediments consist largely of the stable low-magnesium calcite, it is apparent that diagenetic changes should, in general, be earlier initiated and more pronounced in carbonate deposits of shallow water type.

In addition to mineralogical analysis of the Florida carbonate sediments, a determination was made of the percentage of $MgCO_3$ present in the high-magnesium calcite end member. These results are shown in figure 4. The frequency distribution is essentially symmetrical with a sharp peak at about 12.5 percent. This value corresponds very closely with a value of 12.7 percent obtained on a concentration of foraminiferal tests (benthonic) from a composite sediment sample. Other possible sources of the high-magnesium calcite contribution are: algae, generally with $MgCO_3$ concentrations greater than 15 percent; alcyonarian spicules, with concentrations of 11.2 to 14.1 percent; echinoids, 4.0 to 11.5 percent; decapod crustaceans, approximately 10.8 percent; barnacles, approximately 4.0 percent; and ostracodes which may contain from 8.2 to 10.8 percent high-magnesium calcite (Chave, 1954). It seems probable that within the range of ecological situations sampled almost all of the high-magnesium calcite must be furnished by foraminifera, perhaps with significant contributions of alcyonarian spicules in some areas. Algae are important sources of high-magnesium calcite only on the actual reefs themselves because the algal contribution of this mineral comes mainly from adherent and cemented forms.

No dolomite was found in any of the analyses made of carbonate sediments either by routine analysis or by extra-sensitive techniques used on a few test samples. It is

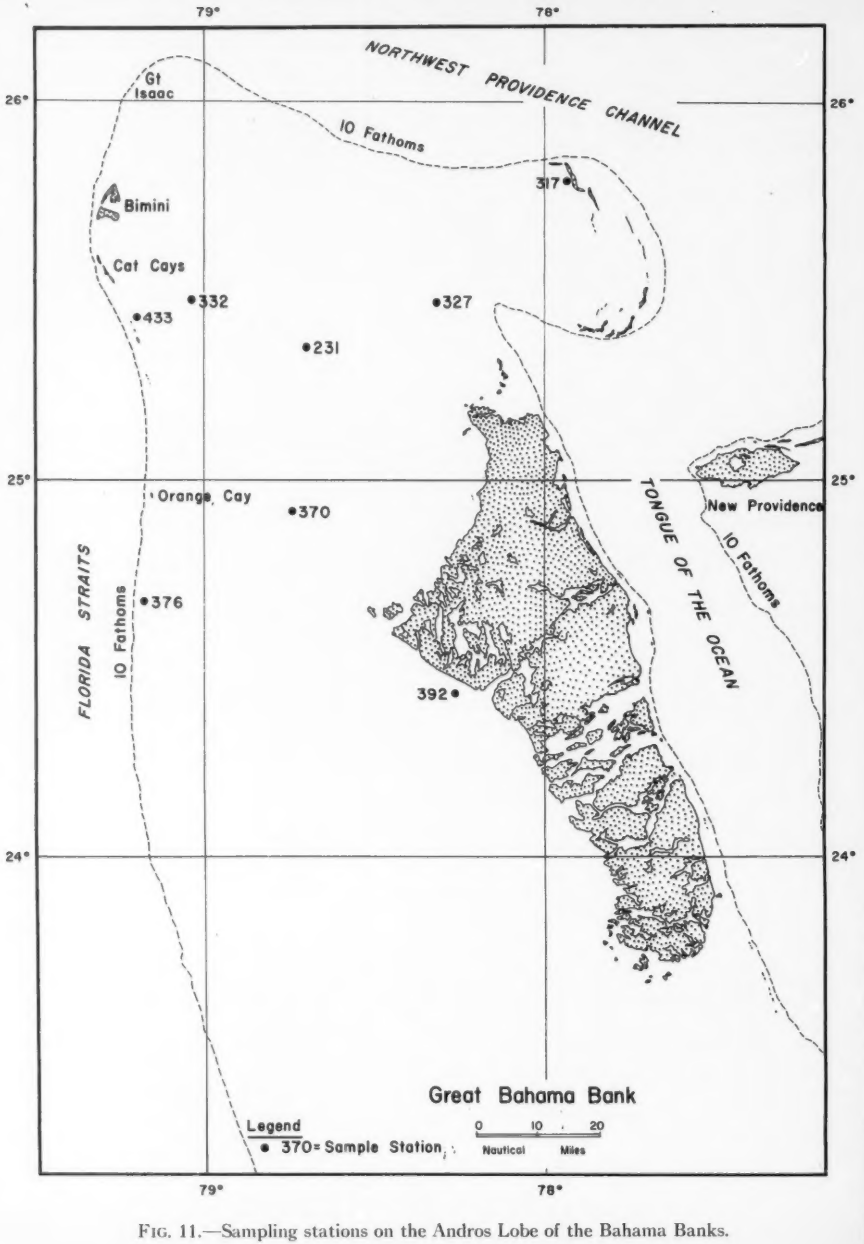

FIG. 11.—Sampling stations on the Andros Lobe of the Bahama Banks.

therefore concluded that neither biological nor physicochemical processes operating in normal marine waters in the areas studied appear to be producing measurable amounts of dolomite.

Mineralogy of Pleistocene Carbonate Rocks

The Pleistocene carbonate rocks were all taken in Florida from the Miami oolite, Key Largo limestone, and Anastasia coquina. These rocks are older than the Sangamon interglacial period, the deposits of which (Pamlico Formation) lie unconformably upon them. The fluctuation of sea level subsequent to the deposition of these carbonates has undoubtedly caused exposure to ground water at least once and may have had important effects on their diagenesis. This complication is essentially universal, however, because of the world-wide nature of Pleistocene sea level fluctuation.

As samples were collected, an effort was made to select specimens in various stages of alteration so that the diagenetic processes could be followed. Examination of figure 5 shows that this objective was only partially reached. In all but two specimens, all of the high-magnesium calcite had already been lost. There is, of course, the possibility that no high-magnesium calcite was originally present. This alternative seems improbable in view of the almost universal presence of high-magnesium calcite in the recent samples from similar environmental situations and the fact that fossils of organisms using high-magnesium calcite in their skeletons can be seen in the rock. It seems reasonable to conclude that high-magnesium calcite is the least stable of the carbonate minerals of sediments and usually recrystallizes quickly. Most of the rock samples analyzed still contained significant amounts of aragonite, ranging from 90 percent in one specimen to an average near 25 percent. Four of the samples no longer contained any measurable amount of aragonite. A general stability sequence for the $CaCO_3$ minerals composing carbonate sediments under natural conditions appears to be as follows: low-magnesium calcite > aragonite > high-magnesium calcite. Of the three original components, only low-magnesium calcite appears to persist for long periods of time, a conclusion previously reached by Chave (1954).

From the evidence discussed above it seems that the diagnetic history of carbonate sediments, under the conditions to which these samples have been exposed, can be traced on the triangular diagram. A sample of normal composition first tends to move to the left side of the triangular diagram through the loss and conversion of its high-magnesium calcite fraction. Secondly, it tends to progress upward along the triangle's left side toward the low-magnesium calcite end-point by the loss and conversion of its aragonite content. The result of early diagenesis is a rock consisting only of low-magnesium calcite.

Volume changes of possible importance in the loss of original porosity and permeability take place in carbonates undergoing diagenesis. The true importance of these changes can only be established by experimental work; however, the conversion of aragonite with a density of about 2.94 gm/cc to low-magnesium calcite with a density of about 2.71 gm/cc will result in a volume increase of approximately 8 percent. For a sediment of the general composition shown in figure 1, a volume increase of 5 to 6 percent would be expected. If the mineral transformation takes place in a body of rock of constant external dimensions, a very important reduction In both porosity and permeability would be expected. Effects caused by converison of high-magnesium calcite to low-magnesium calcite are less important—involving a volume change of only about 1 percent for a sediment of average composition. The actual changes which may take place in the diagenesis of a sediment are complex and depend on the sequence of mineral conversion, the rate of conversion, and whether or not a materials balance is maintained.

It appears probable, however, despite the complications that may become involved, that original porosity is most likely to be lost in the diagenesis of a sediment composed wholly of aragonite and most likely to be retained in one composed of low-magnesium calcite.

Trace Elements

The concentration of the four trace elements most commonly substituting for calcium in the calcium carbonate lattice has been measured on the Florida sediment and

rock samples. In order of decreasing abundance, these elements are: (1) magnesium, (2) strontium, (3) manganese, and (4) barium. The results for both the sediments and the rocks are plotted on a single graph for each element, and in every case diagenesis appears to have resulted in a marked decrease in trace element concentration.

Of the four trace elements only magnesium is sufficiently abundant in sediments to be of major importance. Mineralogical and trace element analyses show that the bulk of this magnesium is incorporated in high-magnesium calcite, a lesser amount in low-magnesium calcite and only small quantities in aragonite. Trace element study shows that upon the conversion of high- to low-magnesium calcite, most of the magnesium is lost from the system. Figure 6 indicates that early diagenesis drops the magnesium level from around 12,500 ppm in sediments to about 1000 ppm for rocks. The magnesium released by conversion of high-magnesium calcite may ultimately contribute to dolomite formation, but our study suggests that it does not generally do so in the rocks in which it originates.

CONCLUSIONS

The following tentative conclusions are reached from this investigation:

1.—Carbonate sediments produced in warm shallow water, either by biochemical activity or a combination of biochemical and physicochemical processes, show a well-defined composition in which aragonite predominates and in which high-magnesium calcite is dominant over low-magnesium calcite. About 70 percent of the average shallow water carbonate sediment consists of unstable forms of $CaCO_3$.

2.—The results of the few analyses made indicate that carbonate sediments indigenous to deep water are composed predominantly of low-magnesium calcite, and that high-magnesium calcite predominates over aragonite. Deep-water carbonates, therefore, are more stable mineralogically than those of shallow-water environments.

3.—Neither biological nor physicochemical processes operating in normal marine waters in the areas studied appear to be producing measurable amounts of dolomite at the present time.

4.—A general stability sequence for the $CaCO_3$ minerals composing carbonate sediments under natural conditions appears to be as follows: low-magnesium calcite > aragonite > high-magnesium calcite. Of the three original components, only low-magnesium calcite appears to persist for long periods of time

5.—Volume changes of possible importance in the loss of original porosity and permeability take place in carbonates undergoing diagenesis. The true importance of these changes can only be established by experimental work.

6.—The $MgCO_3$ content of high-magnesium calcite in the modern sediments studied showed a mode of 12.5 percent, and this constitutes a significant source of magnesium for release during diagenesis. However, this magnesium, at least in the rocks studied, does not appear to form dolomite place although it may later do so at some other place.

7.—Diagenesis of carbonate sediments results in a large loss of trace impurities, including magnesium, from the system so that a materials balance is not maintained.

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FIGS. 1–14, SEPTEMBER, 1961

PETROGRAPHIC FEATURES OF SANDSTONES THAT AFFECT THEIR SUITABILITY FOR ROAD MATERIAL[1]

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ABSTRACT

This study constitutes a portion of a research program to determine the suitability of West Virginia sandstones for use as base or subbase road construction.

Seventy-three thin sections were examined with a petrographic microscope to determine important petrographic features. Fifty hand samples were tested for porosity and permeability.

The relationships between petrographic features and physical tests were studied. The physical tests, conducted by the Department of Tests of the West Virginia State Road Commission, included the percent wear by the Los Angeles test, a freeze and thaw test, a repeated load test, a proposed British test, and a compressive strength test.

Several petrographic features are important in determining the behavior of sandstones in the Los Angeles abrasion test. In argillaceous sandstones, the degree of compaction is important, whereas the degree of cementation and pressure solution are the principal factors determining the percent wear of high-silica sandstones. In calcareous and ferruginous sandstones the degree of cementation is important in determining their strength. The degree of compaction, cementation, and pressure solution of sandstones is reflected by the quantity of large interconnected pores.

The results of the other physical tests indicate that high-silica sandstones would be more suitable than argillaceous sandstones for base and subbase course use under flexible pavement. Differences in compaction in the wet and dry state and breakdown caused by freezing and thawing make argillaceous sandstone undesirable.

INTRODUCTION

The State Road Commission of West Virginia in cooperation with the West Virginia University Engineering Experiment Station is currently conducting a research program to determine the suitability of West Virginia sandstones for use as base or subbase courses in the construction of flexible pavements.

The omnipresence of sandstone in West Virginia and the paucity of economic deposits of limestone in large areas of the State accentuate the importance of utilizing natural sandstone.

At present, the criterion for determining the acceptability of a sandstone for base course use is its percent wear as determined by the Los Angeles Abrasion test.

The purpose of the present study is twofold. In the first phase of the study, the relationship between the petrographic features of sandstones and their behavior in the Los Angeles test was determined. The second objective of the investigation was to determine the relationship between petrographic features and the results of other empirical tests. The empirical tests, conducted by the Department of Tests of the West Virginia State Road Commission, include a freezing and thawing test, a repeated load test, a compressive strength test, and a proposed British test.

Seventy-three West Virginia sandstone samples were collected ranging in age from lower Silurian to uppermost Paleozoic (Dunkard Series). From these, 73 thin sections have been prepared for petrographic analysis. Thirty-two are of high-silica sandstones, 33 are of argillaceous sandstones, and 8 are of calcareous and ferruginous sandstones.

PETROGRAPHIC FEATURES OF THE SANDSTONES

Argillaceous Material

As used in this report, argillaceous material refers to clay size material including the clay minerals sericite, illite, and chlorite; extremely altered feldspar and rock frag-

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ments; and other obscure fine-grained particles. Several types of clay minerals are generally present, but in most cases the different types could not be identified under the microscope. Kaolinite is also a clay mineral but is listed separately in tables 1, 2, and 3 for reasons which will be explained later.

Argillaceous material occurs in various amounts in the sandstones studied. The term "high-silica sandstones" is here arbitrarily assigned to those sandstones containing from 80 to 100 percent quartz, so in these sandstones the amount of argillaceous material is less than 20 percent. Of the high-silica sandstones studied, the clay or argillaceous content ranges from 17 percent (sample 46) to less than 1 percent (samples 37, 45, 48, 63, and 70). The argillaceous content of the other sandstones studied ranges from 9 percent (sample 5) to 37 percent (sample 73).

The distribution of the clay binder or matrix in the high-silica sandstones was studied in detail. It is thought by the writers that the amount of clay in contact with quartz grains has some relationship to the strength of the rock. Clay distributed as a thin layer between quartz grains should produce a less brittle sandstone than one containing the same amount of clay distributed in larger bodies in pores or along bedding planes. Point counts were taken on 600 grains of each high-silica sandstone, and the proportion of each grain in contact with clay, other quartz and guartzite, and kaolinite was noted. The results are shown in table 1. This method is, of course, semiquantitative but does give a good indication of the distribution of argillaceous material in high-silica sandstones.

Kaolinite

The amount of kaolinite was plotted as a separate integer since in most of the samples it is derived from the alteration of feldspar grains and is physically different from the other clay minerals. Kaolinite is distributed as grain-sized patches occupying former feldspar positions. It ranges in abundance from 0 to 7 percent in the high-silica sandstones and from 0 to 10 percent in the other sandstones studied. Megascopically the kaolinite appears as white masses about the size of the other grains. The kaolinite grains are easily plucked and apparently are not bonded to the surrounding quartz grains. The presence of kaolinite in high-silica sandstones of Pottsville age in southwestern West Virginia has been noted frequently by the writer.

Secondary Silica Cement

A study of the secondary silica cement is necessary for a better understanding of the behavior of high-silica sandstones under various strength tests. Criteria considered were the amount of measurable quartz cement, the nature of the cement, and the relationship of cement to pore space, pressure solution, and clay content. The source of the silica cement is not considered.

Generally the amount of secondary cement is inversely related to the clay content (Siever, 1959, p. 69). The amount of secondary quartz measurable, however, is dependent upon the number of original detrital outlines, since secondary silica is deposited in optical continuity with the detrital grain. Thus, if no original detrital outlines are apparent, no cement can be measured, although the actual amount of cement may be considerable.

The degree of cementation is reflected by the nature of the secondary quartz. Incomplete cementation is characterized by prismatic outgrowths of secondary quartz elongated parallel to the "C" crystallographic axis of the detrital grain. The outgrowths do not bridge pore space (fig. 1). Secondary quartz overgrowths enveloping the detrital grain usually bridge voids and indicate more complete cementation (fig. 2). The quantity of secondary quartz measured is not necessarily a reflection of the total amount of secondary silica present, and the measurements are of little value.

The relationship between cement, pore space, and pressure solution was also noted. The amount of pore reduction is increased with an increase in the quantity of secondary silica; however, the amount of pore reduction is also dependent upon the amount and compaction of interstitial material, the degree of pressure solution, and, of course, the amount of original pore space. Original pore space is not considered here, since only the present state of the sandstone is relevant

TABLE 1.—*Petrographic features of high-silica sandstones*

		Composition									Grain* Size	Type Grain Contact					Material in Contact With Quartz			Pore Characteristics		
Sample Number	% Wear	Quartz & Quartzite	Argillaceous Material	Kaolinite	Feldspar	Rock Fragments	Mica	Observed Quartz Cement	Carbonates	Iron Oxides	Grain Size	Flat	Concavo-convex	Slight Interlocking	Moderate Interlocking	Intense Interlocking	Quartz & Quartzite	Argillaceous Material	Kaolinite	Effective Porosity	Thin Section Porosity	Alcohol Drop Test in cm
37	18	97	0.3	0	0	2.3	0	1.3	0	0	f.	31	38	19	9	3	96	4	0	1.1	0.5	2.98
38	34	84	12	0	0.3	0	1	0	1.7	0	v.f.	9	29	61	1	1	51	42	0	9.4	3.0	2.50
39	35	91	8	0	0	0.7	0	4	0	0	v.f.	10	30	22	5	1	74	26	0	8.8	3.0	2.60
40	35	81	13	3	0	0.1	0	5	0	0	f.	10	12	70	6	2	47	44	5	8.5	1.0	2.63
41	36	92	7	0	0.3	0	0.3	0.3	0	0	m.f.	1	4	78	10	7	61	38	0	5.8		2.45
42	38	89	8	2	0	1	0	6	0	0	m.	9	35	40	12	4	47	29	5	9.3	3.5	2.53
43	40	87	12	0	0.7	0	0	3.7	0	0.2	f.v.f.	16	10	64	7	3	44	45	2	6.3	3.0	2.40
44	43	97	2	0	0.3	0	0	2.7	0	0	f.	3	10	72	12	3	74	22	2	10.6		2.27
45	45	98	0.3	2	0	0	0	5	0	0	f.	8	19	66	7	2	83	7	6	8.3		2.30
46	46	82	17	1	0	0	0	1	0	0	f.v.f.	8	28	60	3	1	71	26	1	5.1	3.0	2.37
47	50	89	7	2	0.3	0.3	0	7	0	0	m.f.	5	10	73	8	4	61	33	5	8.2	2.0	2.22
48	52	99	0.3	0	0	0.6	0	0	0	0	m.f.	6	11	76	5	2	95	5	0	4.8	2.0	2.74
49	52	80	16	0.3	0	0	0.3	1	0	2	m.	9	28	57	4	2	43	45	0.4	9.9	1.2	2.30
50	52	95	4	0.7	0	1.3	0	1.7	0	0	congl.	10	27	56	4	4	62	38	2	8.7	4.5	2.05
51	53	84	15	0	0	1	0.2	1.3	0	0.7	m.f.	8	28	63	2	0	44	52	0	9.5	2.0	2.06
52	62	96	3	0.3	0	0	0	1.3	0	1.3	m.f.	15	21	57	4	4	77	21	2	6.8		2.32
53	63	83	12	2	0	0	0.7	3	0	0	m.f.	9	30	59	1	0	56	30	12	7.6	7.0	1.93
54	67	88	8	1	0	1	0	2	0	1.3	f.	8	55	34	2	1	34	48	2	9.8	3.0	2.05
55	72	90	6	2.5	0	0.4	0	2.4	0	0.4	m.	12	31	20	4	1	63	29	2	10.3	6.5	2.0
56	76	91	6	0.4	0	0.5	0	11	0	0	m.f.	13	45	40	0.5	0.2	64	36	0	10.0		2.0
57	77	93	7	0	0	0.5	0.2	5	0	0	f.	11	31	52	4	2	66	34	0	10.7	2.0	2.07
58	79	89	10	0.7	0.3	1.3	0	2	0	0	m.	7	21	66	5	1	59	39	0	11.0	7.5	1.88
59	87	96	2.8	0.4	0	0.2	0	5	0	0.3	m.c.	5	16	75	3	1	94	3	2.5	8.4	6.7	2.20
60	90	96	2	2	0	1	0	6	0	0	m.c.	8	17	70	4	1	44	40	14	9.5	10.0	1.92
61	90	96	5.5	0.5	0	1.1	0	4	0	2	m.	16	20	61	3	0	54	31	1	9.1	3.7	2.14
62	95	96	2	2	0	0	0	8	0	0	m.	12	14	75	6	3	69	33	8	11.0	8.5	1.92
63	98	100	0.3	0	0	0	0	3	0	0	m.	13	39	48	0	0	99	1	0	9.2	10.7	1.95

* v. = very; f. = fine; m. = medium; c. = coarse; congl. = conglomerate.

TABLE 2.—*Petrographic features of argillaceous, calcareous, and ferruginous sandstones*

		Composition								Grain* Size	Compaction	Pore Characteristics		
Sample Number	% Wear	Quartz & Quartzite	Argillaceous Material	Kaolinite	Rock Fragments	Feldspar	Mica	Calcite	Iron Oxides	Grain Size	No. of Touching Grains	Effective Porosity	Thin Section Porosity	Alcohol Drop Test in cm
1	34	43	35	0	0.8	3	8	8	2.0	v.f.	0.84			
2	34	47	34	1.5	1.3	5	10	1.5	0	f.v.f.	1.44			
3	36	46	32	0	1.5	5	10	6.0	0.5	f.v.f.	1.51			
4	37	62	21	6	5	3	3	0	2	m.f.	2.09			
5	39	51	30	1.5	3	4	6	3	2	f.	1.63			
6	40	55	27	0.3	6	5	1.3	4	0	f.	2.09			
7	42	47	29	0.3	1.5	7	6	6	4.0	f.v.f.	1.70			
8	44	52	31	3.5	5	3	5	0	1	f.	1.87			
9	47	56	29	0	0.8	3	5	2	4.5	f.v.f.	1.49			
10	50	58	19	6	4	9	3	1	1	m.m.f.	2.12			
11	56	58	20	5	3	10	1.3	2	1	m.m.f.	2.16			
12	57	52	31	4	3	5	5	0	0.5	m.f.	1.69			
13	57	59	20	8	6	4	3	1	0	m.	1.84			
14	62	62	19	5	8	2	5	0	0	m.f.	1.78			
15	66	79	8	2	3	0.5	0	0	0.5	m.m.c.	2.97			
16	68	71	9	7	6	3	0.8	1	2	m.c.	2.60			
17	77	88	8	1.5	1	1	0	0.5	0	m.m.c.	3.35			
18	81	51	24	2	8	6	5	0	4	m.m.f.	1.87			
19	83	53	21	6	5	6	6	0	1	m.c.	1.79			
20	97	70	11	4	9	3	0.8	0	1.5	c.	2.31			
21	98	59	13	11	8	6	0.5	0	0.5	c.	2.02			
22	100	73	10	7	6	3	1.5	0	0	c.	2.30			
23	36	67	25	0.7	2	0	2	0	0	f.v.f.	3.35	6.6	2.0	2.60
24	38	53	29	5	0.5	2	9	1	0.8	f.v.f.	1.88	3.7	0.7	2.15
25	64	75	27	7	0.2	1	2	0.2	2	f.	1.95	4.2	2.3	2.0
26	70	61	20	4	1	4	4	7	0.4	f.v.f.	3.12	12.6	3.6	1.85
27	93	59	19	10	0	3	9	0	0	f.	2.02	12.3	5	1.75
28	95	62	14	8	0	4	2	6	0.2	m.f.	2.64	14.8	5.3	1.60
29	12	52	17	0	8	1	1	0.2	21	v.f.		2.0	0	2.98
30	17	63	26	0	4	0.8	0	0	7	f.v.f.		4.5	0.1	2.80
31	17	58	12	0	2	2	0	0	25	v.f.		3.0	0	2.86
32	53	55	13	2	4	7	1.5	17	0	m.m.f.	1.93			
33	22	34	10	0	2	5	2	46	1	f.		2.8	0.1	2.70
34	22	36	11	1	2	3	1	46	0	m.f.		6.2	0.1	2.77
35	38	53	22	0	0	6	11	27	0	m.f.		7.3	0.4	2.20
36	38	50	16	0.5	0.8	3	4	21	5	m.f.		7.9	0.7	2.27

* v. = very; f. = fine; m. = medium; c. = coarse.

for comparison with strength tests.

As well as causing a reduction of pore space, pressure solution is associated directly or indirectly with secondary quartz cementation. Pressure solution is defined as solution of quartz grains at points of grain contact. The dissolved silica may be deposited on the same grain at points of no contact, in other low pressure areas, or may be carried out of the system and deposited eleswhere. Heald (1950, p. 632) observed pressure solution of quartz and other grains in certain Paleozoic sandstones of West Virginia. He further suggested that clay between grains plays a

TABLE 3.—*Petrographic features of samples 64-73*

Sample Number	% Wear	Composition								Grain* Size	Compaction	Pore Characteristics		
		Quartz & Quartzite	Argillaceous Material	Kaolinite	Rock Fragments	Calcite	Feldspar	Mica	Observed 2nd Quartz	Grain Size	No. of Touching Grains	Effective Porosity	Thin Section Porosity	Alcohol Drop Test in cm
64	78	94	1	1	3	0	0	0	7	m.	5.18	5.7	3.7	2.35
65	66	92	5	2	1	0	0	0.1	9	m.	4.29	11.9	8.5	2.15
66	85	82	9	7	0.3	1.0	0.3	0.3	1.2	m.	3.2	16.9	6.2	1.85
67	35	60	27	6	0.3	0.7	1	7	1.3	f.	1.58	11.8	2.0	2.15
68	65	62	17	5	1	8	4	4	0.3	f.	2.27	10.0	1.5	2.85
69	38	60	22	4	1	9	2	2	0	f.	1.84	10.0	0.2	2.33
70	54	99	0.3	0.3	0	0	0.3	0	15	m.f.	5.82	9 8	4.2	1.85
71	40	96	3.5	0	0	0	0.2	0	2.5	m.c.	6.23	8.3	6.5	2.37
72	63	72	21	2	0	3	1	2	0	m.f.	2.41	12.6	1.0	2.10
73	63	60	37	2	0	0	1	1	0	f.	1.32	13.2	1.6	1.88

* v. = very; f. = fine; m. = medium; c. = coarse.

role in the development of pressure solution similar to that of a catalyst (Heald, 1956, p. 25; 1959, p. 251). Sutured contacts between quartz grains are indicative of highly pressolved zones. Suturing and the interpenetration of grains because of secondary quartz growths were measured to determine the degree of interlocking of quartz grains. The nature of grain contacts in high-silica sandstones was measured by estimation of the percent of tangential, flat, concavo-convex, or sutured contacts. Grain contacts, as used in this investigation, include those between detrital grains as well as those between secondary quartz overgrowths. No tangential contacts were observed in the high-silica sandstones studied.

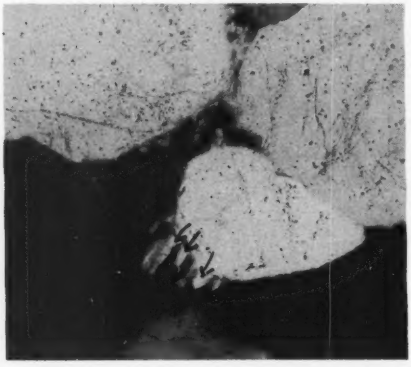

FIG. 1.—Prismatic outgrowths of secondary quartz (arrows) with their long dimension parallel to the c-axis of the detrital quartz grain. Crossed nicols. ×85.

The degree of interpenetration or suturing was further differentiated by measurement of the average ratio of wave length to amplitude of the sutures. Sutured and interpenetrated contacts were classed as slight (wave length/amplitude = $\frac{1}{2}$), moderate (wave length/amplitude = 1), and intense if the interlocking restricted grain movement in two directions (fig. 3). This method is semi-quantitative at best, but does show differences in the degree of interlocking of quartz grains. It is thought that the more intense sutures or interpenetrations increase the cohesiveness of the sandstone. The results of these measurements are shown in table 1.

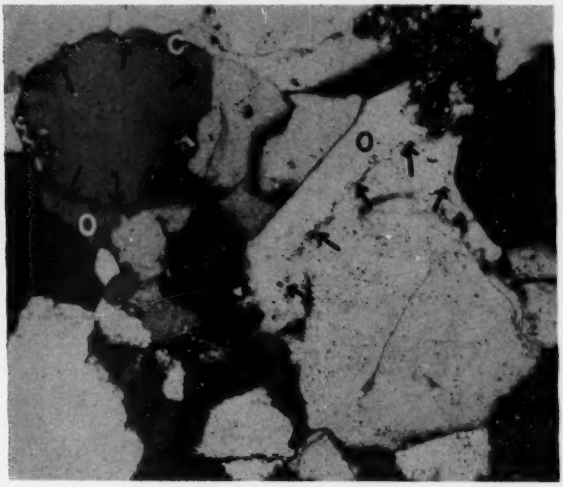

FIG. 2.—Secondary quartz overgrowths "O" and dust rings (arrows) outlining the original detrital grains. Crossed nicols. ×100.

Carbonate Cement

The occurrence of carbonate cement in the high-silica sandstones of the Pottsville Series of West Virginia is uncommon. Of the high-silica sandstones studied, only two contain this type of mineral cement. Sample 38 contains 1.7 percent siderite, and sample 66 (Conemaugh Series) contains 1 percent calcite.

Calcite occurs In various amounts in the other sandstones studied. It is the dominant bonding material in five of the samples studied (fig. 4). Sample 32 contains 17%, sample 36 contains 21%, sample 35 contains 27%, and samples 33 and 34 each contain 46%. In each of the above sandstones calcite is in the form of secondary cement replacing quartz, feldspar, rock fragments, and argillaceous material. This secondary replacement causes good interlocking between cement, grains, and matrix.

Iron Oxide Cement

Iron oxide is the dominant cement in the ferruginous samples 29, 30, and 31 where its percentage is 21, 7, and 25 respectively (fig. 5).

In the high-silica sandstones studied, iron

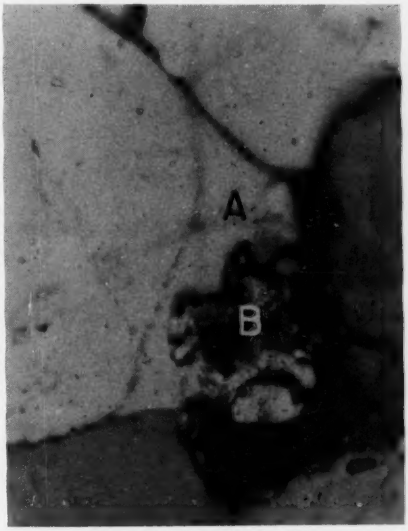

FIG. 3.—Intense interlocking of grains produced by secondary overgrowth "A" engulfing a quartzite fragment "B." Crossed nicols. ×250.

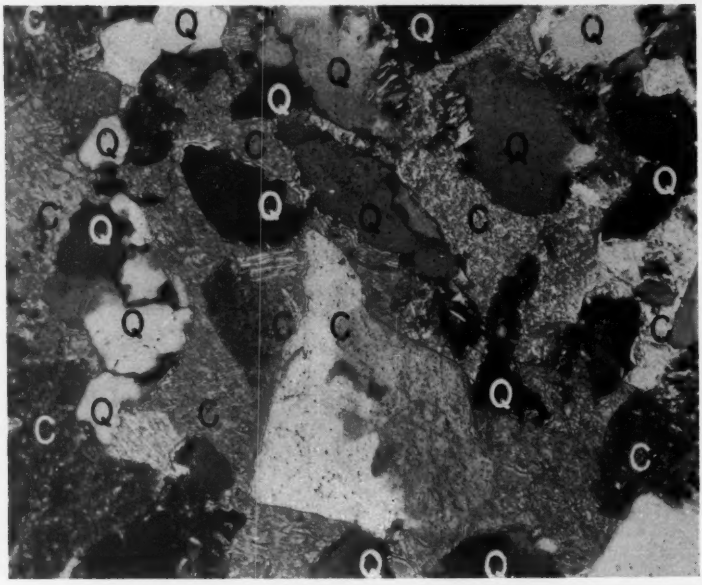

FIG. 4.—Calcite cemented sandstone: quartz grains "Q" are "floating" in calcite cement. Crossed nicols. ×100.

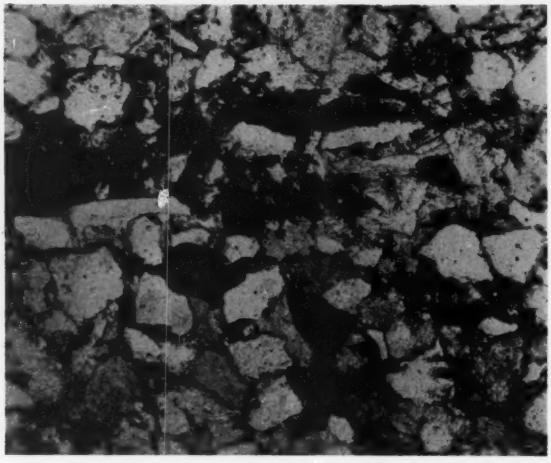

FIG. 5.—Ferruginous sandstone: iron oxide cement (black) is the dominant bonding agent. Plain light. ×100.

oxide is common but occurs only in minor amounts. The sandstones were collected above the water table in a zone susceptible to oxidation. High-silica sandstones are generally brittle and contain numerous joint planes. These joints are structural avenues for passage of supergene (descending) waters, and oxidation is common at and near these planes. The degree of oxidation is dependent upon the porosity and permeability of the sandstone as well as joint fractures. At sample locations 44, 45, and 52, differential oxidation was observed. The more permeable and porous zones were more intensely oxidized than those less porous and permeable. Coincidentally, those zones of little oxidation and low porosity were zones of greater secondary quartz cementation. Several of the samples studied contain small amounts of iron oxides.

The effect of oxidation and weathering on high-silica sandstones can be observed by comparing samples 42 and 47. Sample 42 represents the unoxidized portion of the sandstone and 47 represents the oxidized portion of the same rock.

Pore Space

A study of the pore space in sandstone is an important consideration in determining its strength. Porosity and permeability measurements were taken to determine the amount and distribution of pore space in the sandstones. The sandstones considered (samples 23–63) include high-silica, argillaceous, calcareous, and ferruginous sandstones.

Porosity may be defined as the percent of the total volume of a rock not occupied by solid material. This pore space may be of various dimensions and may or may not be connected. Primary or original porosity is the porosity of the original sediment and is controlled by grain size, grain shape, packing, and the amount of argillaceous material. As sediment is transformed into rock, original porosity is modified by several postdepositional processes: compaction, cementation, intrastratal solution, and weathering or decementation. As a result of these complicated depositional and postdepositional processes, the porosity of a bulk sample of sandstone may or may not be uniform. Thus the measurement of porosity is difficult, and some of the measurements are not representative. Several methods were employed.

Effective porosity.—The results of effective porosity measurements by the Stevens porosimeter were considered the most reliable of the porosity tests run. The Stevens porosimeter measures the volume of air removed from the pores of a cylindrical sample ½ inch in diameter and 1 inch in length. The volume of the sample is measured by a pycnometer, and the determinative formula is:

$$\text{Effective porosity} = \frac{\text{Volume of pores}}{\text{Volume of sample}}$$

The porosimeter method is somewhat better than methods employing a fluid medium but is not ideal for purposes of this study. The necessity for cutting a cylinder to obtain consistent results may add an element of error. Cylindrical samples were cut using a half-inch diamond core drill on a drill press under a flow of water. This cutting operation could, in friable sandstones at least, induce porosity. The more friable sandstones studied were subjected to other effective porosity determinations. Approximately 40 pounds of each sandstone was collected for strength tests. The size of the sample used in the Stevens porosimeter test is approximately 30 gm. The questionable representative value of such a small sample and the time consumption of the test are other objectionable factors. The results of this test are listed in tables 1, 2, and 3.

Porosity by thin section examination.—Porosity determinations were made by examination of thin sections under the petrographic microscope. In figure 6 the percent of the effective porosity, which is submicroscopic, is plotted against percent interstitial material. This figure shows that there is an increase in submicroscopic pore space with an increase in interstitial material.

Permeability.—Permeability was determined for several samples of which suitable cores were cut. The permeability values of the samples tested are low since obviously weak and permeable rocks would not be sampled for road use.

Problems encountered in permeability measurements were similar to those associated with effective porosity determinations

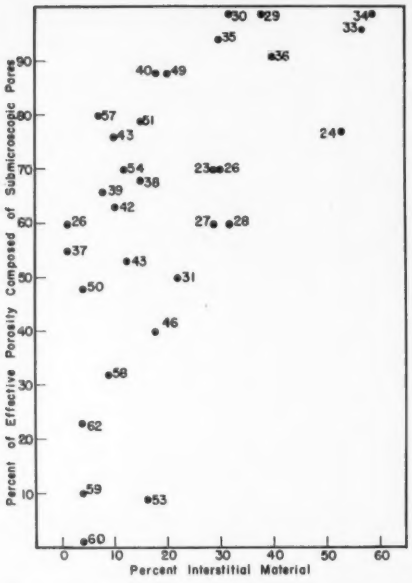

FIG. 6.—Relationship between submicroscopic pores and percent interstitial material.

by the Stevens porosimeter method. It was noted that much of the variation in permeability was caused by the presence of such megascopic structures as bedding planes and joint fractures. Samples 30 and 31 both have low porosity and are very similar microscopically; however, sample 31 is notably more permeable. This difference is caused by a prominent bedding plane passing through sample 31.

Pore size determinations by the alcohol drop test.—A semiquantitative test is presently used by the West Virginia Highway Aggregate Research Project for quick determination of the probable percent wear of a sandstone. This empirical test consists of deducing the absorption rate of a sandstone by measuring the diameter of an expanded drop of alcohol on a surface perpendicular to bedding. The surface is held in a horizontal position, so that the absorption rate in a direction parallel to bedding may be ascertained. The principle, in simplest form, is that the larger the expanded drop of alcohol on the surface of the sandstone, the less the absorption rate in a direction parallel to bedding.

The main advantage of this test is that a large bulk sample may be tested quickly, and any variation in absorption is easily seen. The average of several measurements is then representative of a bulk sample.

Samples 23 to 63 each constitute approximately 40 pounds of sandstone and were collected for precent wear determination only. Each of these bulk samples was subjected to the alcohol drop test, and the average results are recorded in tables 1, 2, and 3. A small diameter indicates rapid absorption, and a large diameter indicates slow absorption. The method is valid only when the diameter of the expanded drop of alcohol on the sandstone is between 1 and 3 cm. In rocks of small permeability, the alcohol drop beads and tends to run off the surface, and even in the most porous and permeable rocks the diameter of the expanded alcohol drop is never smaller than 1 cm in diameter.

A comparison of the results of the alcohol drop test with those of the Stevens porosimeter (fig. 7) are interesting. The samples with a low absorption rate (large diameter of expanded drop) per given effective porosity

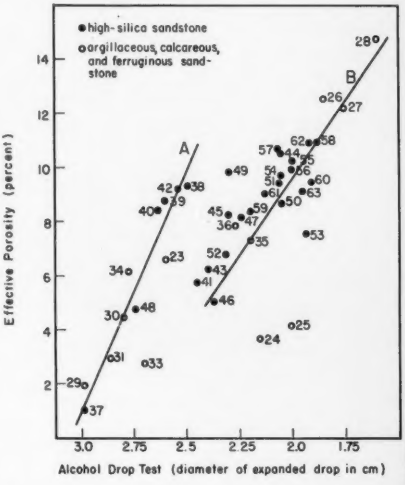

FIG. 7.—Relationship between the alcohol drop test and effective porosity.

are argillaceous, calcareous, and ferruginous sandstones that are fresh and highly compacted and high-silica sandstones that have been well cemented or pressolved. These samples are plotted near curve A in figure 7. The samples with a high absorption rate per given effective porosity are argillaceous sandstones that have been oxidized, weathered, or contain clay minerals altered from feldspar, and high-silica sandstones that have been oxidized and weathered or contain large open pores because of incomplete cementation and lack of pressure solution. These are near curve B in figure 7. The samples between the two curves contain features intermediate between the two groups.

The low absorption per effective porosity in sandstones near curve A is caused by small pore spaces and lack of free interconnections between the pores.

The higher absorption rate in the sandstones near curve B is caused by larger pores or more interconnected pores which are primary or result from weathering and alteration. The interconnecting pores are not entirely obvious from thin section examination, since a large percentage of the pore space of argillaceous sandstone is submicroscopic and the exact size of the pores is not ascertainable. A consideration of liquid flow through a porous medium supports this contention.

The action of a single phase liquid in the pore space of a rock is dependent upon two forces, gravity and capillarity (Meinzer, 1923, p. 18). The action of capillarity varies inversely with the diameter of the pores, and the converse is true of the effect of gravity. In the alcohol drop test a drop of alcohol is placed on the surface of a sandstone and gravity is effective in a downward direction parallel to bedding. If the pores are large, gravity tends to move the alcohol downward; however, if the pores are very small, the downward migration of the alcohol is restrained, and the lateral spreading becomes more effective. Thus, it seems logical that the alcohol drop test gives an indication of the quantity of the larger interconnected pore space of a sandstone. The pore size which constitutes the lower limit of the effect of gravity over capillarity is submicroscopic and not easily ascertained.

Grain Size

The average grain size of each sample was estimated under a binocular microscope by comparison with grains of known size. The method is, once again, semiquantitative, but, since some of the sandstones contain quartz cement, sieve analysis was impractical. The determinative results are listed in tables 1, 2, and 3.

Compaction of Interstitial Material

In the argillaceous sandstones studied (table 2), the grains consist of quartz, quartzite, feldspar, metamorphic rock fragments, mica, very small amounts of chert, and accessory minerals. The interstitial material consists of the argillaceous minerals illite, chlorite, and sericite and small amounts of the chemically precipitated cements calcite and limonite. The compaction of interstitial material is an important attribute affecting the strength of argillaceous sandstone. Petrographic features which influence or are related to compaction are grain size, amount of interstitial material, the number of touching quartz grains, mineral alteration and weathering, and the number of large and interconnecting pores.

The grain size and amount of interstitial material are related in a manner shown by figure 8. There is an increase in interstitial material with an increase in grain size.

Enough interstitial material is present in the fine-grained sandstones to completely surround most of the grains and prevent them from touching. The more interstitial matter in a sandstone, the fewer number of touching grains (fig. 9).

The sandstones with smaller amounts of interstitial material contain more grains and have more touching grains. When these rocks undergo compaction, the bridging effect of the touching grains prevents the interstitial matter from receiving the strong pressures required for tight compaction.

Expansion caused by alteration of minerals and weathering causes a loosening of interstitial material. This loosening is accompanied by an increase in large interconnected pore space. The quantity of large and interconnected pores, therefore, gives an indication of the amount of compaction of interstitial material.

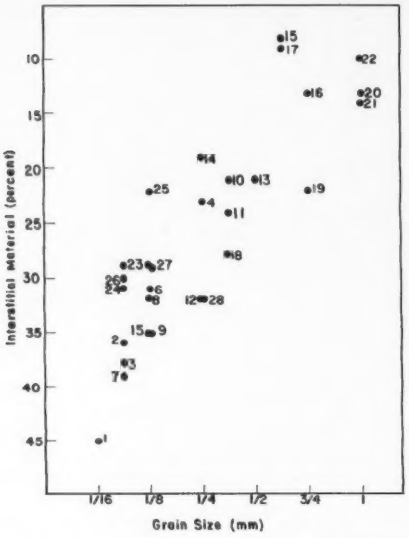

FIG. 8.—Relationship between grain size and interstitial material in argillaceous sandstones.

RELATIONSHIP BETWEEN INHERENT FEATURES OF SANDSTONE AND THE LOS ANGELES TEST

To better understand the features of a sandstone that affect its percent wear in the Los Angeles abrasion test, the procedure for running the test should be evaluated.

Approximately 40 pounds of rock are sampled for test purposes. Sampling is usually done with a sledge hammer, and the rock is broken down into pieces weighing approximately 2 pounds or less. Experience has shown that in this initial "breaking up" of the rock, the planes of easiest fracture are usually along planes of inherent weakness such as bedding planes, joint fractures, and uncemented zones.

The 40 pounds of broken rock is then placed in a cone crusher which further breaks the rock into smaller fragments (1 in and less in diameter). This crushing action again breaks the rock along inherent planes of weakness, so a thin bedded sandstone is broken into tabular fragments and a massive sandstone into equidimensional fragments.

The crushed fragments are then screened, and 5000 gm (size less than $\frac{3}{4}$ in to greater than $\frac{3}{8}$ in) are used for testing. Thus, the wear of particles greater than $\frac{3}{4}$ inch in diameter that contain planes of weakness is not truly tested.

Five thousand grams of material is then placed in the Los Angeles machine, a steel drum 28 inches in diameter. The drum completes 500 revolutions in approximately 15 minutes. Each revolution brings the sample and 11 steel balls to the top of the machine where both fall to the bottom with a subsequent impact of a ball against sandstone. The result of this impact separates any remaining minute planes of weakness, and furthermore, tends to disaggregate the fragment by breaking the bond between the sand grains. If this bond is extremely great, as in a quartzite, the bond between the grains is as strong as the grain itself, and fracture occurs through the grains.

The contents of the machine are removed and screened on numbers 4, 8, and 12 mesh screens. The material retained above the number 12 mesh screen is weighed. This

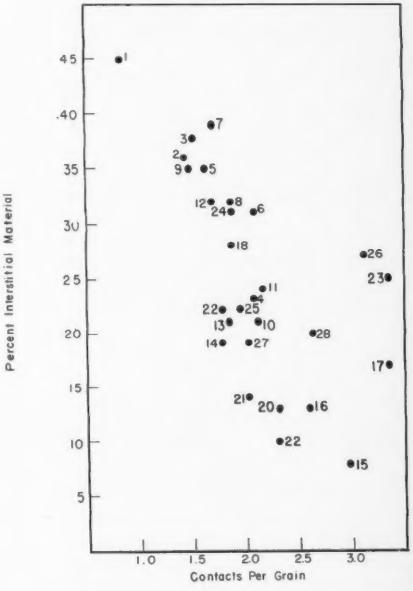

FIG. 9.—Relationship between the number of touching quartz grains and interstitial material in argillaceous sandstones.

weight is subtracted from the original 5000 gm, and the remainder is multiplied by 2 and marked off two decimal places to the right. The result constitutes the percent wear of the sandstone. If the percent wear is less than 64, the stone is accepted for subbase material.

Two points of interest should be noted here. First, those sandstones which are thin bedded (usually argillaceous) are broken along bedding planes in the abrasion machine, and the shape of the fragments which are screened are tabular or platy. The material may be broken down considerably, but the long dimension of the platy fragments allows them to be retained above the number 12 screen. Thus a shale would show a lower percent wear than a nonlayered rock of the same strength. The second point is that a sample may be completely disaggregated to grain size, but, if the grain size is larger than number 12 mesh, (1.55 mm in diameter), it will have a low percent wear. Sample number 50 is of this nature, and an examination of the final screenings showed complete disaggregation to grain size, yet the wear according to the Los Angeles test was only 52 percent.

The test has two drawbacks: (1) planes of weakness are not truly tested, especially in particles above $\frac{3}{4}$ in, and (2) the strength of extremely coarse-grained sandstones is not represented. The test does indicate the ability of the sand grains (less than 1.55 mm in diameter) to hold together under impact; in other words, it measures the coherence between the sand grains.

Any comparison of the inherent characteristics of a sandstone with its percent wear must, therefore, take into consideration the factors which affect its coherence.

The grains of a sandstone may be held together by various types of bonding material. This bonding material may consist of argillaceous minerals, calcite (and other carbonates), silica, and iron oxides. For comparison purposes, the following examples of road material of various types which have a high resistance to the Los Angeles abrasion test are noted: solid vein quartz has a percent wear of 21; a sample of dense calcite-cemented sandstone (Ridgeley Sandstone) tested 12% wear; a dense limestone tested 11% wear; and a ferruginous sandstone had 12% wear. Each of the above samples is nonporous.

Pore Size and the Los Angeles Test

Of the pore studies conducted, the results of the alcohol drop test (fig. 10) are most significant. This test gives an indication of the quantity of large and interconnecting pores in a sandstone. The curve (fig. 10) demonstrates the inverse relationship between percent wear and the quantity of large pore space.

Pore size in high-silica sandstones.—The coherence of the high-silica sandstones collected is caused by silica cement and in most cases small amounts of argillaceous material as well as interlocking resulting from pressure solution. The degree of cementation and pressure solution and compaction of interstitial material is related to pore space, as these processes tend to reduce porosity.

Most of the high-silica sandstones with many large pores have a high percent wear,

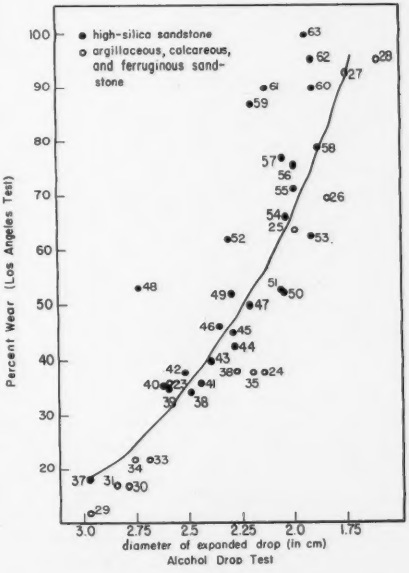

FIG. 10.—Relationship between the alcohol drop test and percent wear (Los Angeles test) of high-silica, argillaceous, calcareous, and ferruginous sandstones.

whereas those with fewer large pores have a low percent wear (fig. 10). It is obvious from an examination of figure 10 that there are exceptions to this general relationship. A few sandstones with very small amounts of argillaceous material (less than 5 percent) are weaker than others containing the same quantity of large pore space and consequently are to the left of the curve (nos. 48, 52, 59, and 61).

These weaker sandstones have secondary quartz cement which is partially in the form of prismatic outgrowths into pore space (fig. 2) and generally indicate incomplete cementation. The outgrowths occupy former pore space, hence reduce porosity, but do not interlock or bridge the voids. Essentially the situation shows reduction of pore space without subsequent increase in cohesion, and the grain-to-pore surface area is relatively high. This explains why these sandstones are weaker than other sandstones with the same number of large pores.

Pore size in argillaceous, calcareous, and ferruginous sandstones.—For comparison purposes, the alcohol drop test was also applied to argillaceous, calcareous, and ferruginous sandstones. The results show that these sandstones are slightly to the right of the general curve for high-silica sandstones (fig. 10). This suggests that for a given quantity of large pore space (same diameter of expanded drop) high-silica sandstones are slightly weaker under the Los Angeles test; however, the differences are so small that they may not be significant.

The degree of compaction of interstitial material in argillaceous sandstones is indicated by the alcohol drop test, since compaction is responsible for a reduction in the amount and size of pore space. The possibility of high compaction is enhanced by a high clay content since the number of touching grains is small. This resulting compaction, however, may be partially or wholly destroyed by later expansion caused by alteration of clay minerals or weathering. Expansion accompanied by an increase in porosity will cause a loss in strength. Thus a highly argillaceous sandstone may have a high percent wear.

The degree of cementation in calcareous and ferruginous sandstones is also indicated by the number of large and interconnecting pore spaces. A large quantity of calcareous or ferruginous cement causes a reduction in porosity as well as an increase in the interlocking bond between cement and grains. These conditions cause a low percent wear.

Effective Porosity and the Los Angeles Test

The direct relationship between effective porosity and percent wear is only very general. Close correlation was not expected since the porosimeter method is a measurement of both large and small pores. The possibility of inducing porosity by sample preparation and the questionable representative value of a single small test sample would tend to mask any real relationship between effective porosity and percent wear.

Permeability and the Los Angeles Test

Permeability of the samples measured does not correlate with the results of the Los Angeles test. Some permeability measurements were high because of bedding planes, joint planes, or other fractures. Since these features are not controlling factors in the Los Angeles test, permeability bears little relationship to the results of this test.

Kaolinite, Porosity, and the Los Angeles Test

Figure 11 kaolinite content plus effective porosity of high-silica sandstones plotted against percent wear. It is believed by the writers that the kaolinite was derived from alteration of feldspar and offers negligible grain support. This kaolinite can be considered to have essentially the same effect as pore space with respect to grain coherence. Porosity plus kaolinite plotted against percent wear show a closer correlation than porosity alone plotted against percent wear. This relationship holds true only for high-silica sandstones, since argillaceous sandstones contain many relatively small pores, and effective porosity of small pores bears little relationship to strength. The difficulty of obtaining representative porosity measurements by the porosimeter, however, is one factor which tends to obscure this relationship.

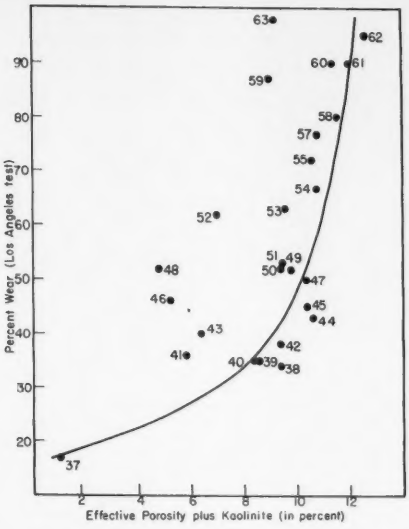

FIG. 11.—Percent kaolinite plus effective porosity of high-silica sandstones plotted against percent wear (Los Angeles test).

Illite derived from alteration of feldspar probably affects the strength of high-silica sandstones in a manner similar to that of kaolinite. This type of illite was observed in samples 43 and 46. Both samples occur to the left of the curve in figure 11.

Grain Size and the Los Angeles Test

There is no direct relationship between average grain size and percent wear. Generally fine-grained sediments contain more argillaceous material than coarse-grained sediments, and would be expected to undergo greater compaction and have less porosity. However, cementation, intrastratal solution, and weathering tend to modify this relationship, especially in high-silica sandstones, so exceptions are almost as numerous as the rule. For example, it is noted that some medium- or medium-coarse, high-silica sandstones (samples 39, 71, and 49) have a lower percent wear than some of the fine-grained, high-silica sandstones (samples 57 and 54). Large pores have been reduced by cementation and pressure solution to a greater extent in those sandstones with a lower percent wear. It is also noted that some very fine- to medium fine-grained argillaceous sandstones (samples 31, 27, and 28) have a high percent wear. Porosity has been increased in these sandstones by weathering or other causes. They contain larger pore space than other argillaceous sandstones with the same quantity of interstitial material (fig. 6).

Effect of Weathering on the Los Angeles Test

Intense weathering obviously causes a decrease in the strength of a rock; however, the effect of slight weathering is not known. Several of the sandstones examined contain iron oxides, but in only one sample was the unoxidized portion of the sandstone available for comparison. Sample 42 represents the unoxidized portion of a high-silica sandstone and 47 represents the oxidized portion of the same rock. The percent wear of the unoxidized and the oxidized portion of the rock was 38 and 50 respectively. The alcohol drop test indicates an increase of interconnecting pore space in the oxidized portion of the rock, and thin section examination shows that the argillaceous fraction is iron stained, and some pores are filled with iron oxides. The increase in interconnecting pore space is caused by other weathering factors associated with oxidation.

PETROGRAPHIC FEATURES AND EMPIRICAL TESTS

Samples 64 to 73 were collected for the purpose of comparing argillaceous and high-silica sandstones under various empirical tests. These tests include a freezing and thawing test, a repeated load (fatigue) test, and a proposed British test. Petrographic features of the 10 samples are shown in table 3. There are 5 argillaceous and 5 high-silica sandstones.

Samples 1 to 28 are argillaceous sandstones which were subjected to compressive strength tests. The above tests were conducted by O. J. Hahn[3] for the Department of Tests of the State Road Commission of West Virginia.

[3] Research Assistant, West Virginia University.

Freezing and Thawing Test

A freezing and thawing test was conducted to determine the relative strengths of the 10 sandstones under repeated freezing and thawing cycles. Particles of definite sizes were tested and the amount of breakdown recorded at the end of each cycle. For comparison, the slope of degradation for particle sizes retained between ¼-in and the #4 screen at the end of each cycle is plotted against the number of freeze and thaw cycles. A high slope indicates a rapid breakdown. The results are as follows:

Sample No.	Slope	% Clay
64	6.0	1.0
65	6.5	5.0
66	10.0	9.0
71	16.5	3.5
67	19.5	27.0
70	20.5	0.3
69	33.0	22.0
68	48.0	17.0
72	175.0	21.0
73	195.0	37.0

Only a general relationship between petrographic features and the freezing and thawing test can be ascertained. It is obvious that the high-silica sandstones tested are more stable than the argillaceous sandstones. Resistance to freezing and thawing is not, however, directly a function of the amount of clay present or the total amount of quartz. Sample 70 contains the least clay (0.3 percent) and the most quartz, but is comparable to sample 67 which has 27 percent clay. There is also no direct correlation between percent wear by the Los Angeles test and the freezing and thawing test.

Factors of importance appear to be the nature of the pore space and the amount and type of argillaceous material. The freezing action of water in sandstone might be compared to the wetting and drying of sandstone. According to Lewis and others, (1953, p. 953), the movement of water into the pores of a rock depends on gravity and capillarity. Water moves in and out of the pores of a rock much more rapidly if the pores are large and capillarity is negligible. Water is easily evaporated from large pores but is retained in smaller pores by capillarity.

Of the 10 sandstones under discussion, the high-silica sandstones contain large pores and the argillaceous sandstones contain small pores. It is suggested that the freezing water is able to expand through the larger pores of the high-silica sandstones without exerting a breaking stress on the quartz grains. In argillaceous sandstones the force of capillarity retains some of the water, and expansion during freezing may be effective in breaking down the clay "binder."

The type of bonding material apparently has some significance. Sample 67 (27 percent clay) is more stable than the other argillaceous sandstones. This rock differs from the other argillaceous sandstones with respect to mica content and the type of argillaceous material. Mica constitutes 7 percent of this sandstone. The argillaceous material is unique, in that it includes a large percentage of sericite (fine-grained mica). The exact proportion of sericite to other argillaceous material is not ascertainable by thin section analysis, although a relatively high percentage is apparent. A detailed study of the types of argillaceous material is not within the scope of this study, but additional data on argillaceous material might be helpful in determining behavior under freezing and thawing tests.

Compressive Strength Test

The compressive strength of argillaceous sandstones (samples 1–23) in pounds per square inch was measured both perpendicular and parallel to the bedding planes. There appears to be very little relationship between the compressive strength and the percent wear except in the sandstones with low compressive strengths. These sandstones have very high percent wear.

Wolkodoff, (1955, p. 1683), has shown that the compressive strength of high-silica sandstones with good grain-to-grain articulation is greater than that of sandstones characterized by point-to-point contacts and incompetent interstitial material (clay).

Results of compressive strength tests on solid stone indicate that the sandstone is much stronger than would be required for traffic conditions.

Repeated Load (Fatigue) Test

This test measures the resistance of

crushed stone under a constant given load applied one million times. The test was applied to samples 64 to 73.

The results of the fatigue test on crushed stone show striking differences between argillaceous and high-silica sandstones. Figures 12 and 13 show the family of curves resulting from testing the samples in both the wet and dry state. The horizontal coordinate represents the load applied in pounds per square inch, and the vertical coordinate represents the slope (inches per cycle) of the curve produced when the penetration of the vibrating head into the cylinder containing the sample is plotted against the number of cycles per load.

An examination of figures 12 and 13 reveals two trends. The argillaceous sandstones are stronger than the high-silica sandstones in the dry state, but, in the wet state, the argillaceous sandstones are weaker. The strength of the high-silica sandstones is relatively unchanged in the wet state.

High-silica sandstones having a greater number of quartz grains in contact are more

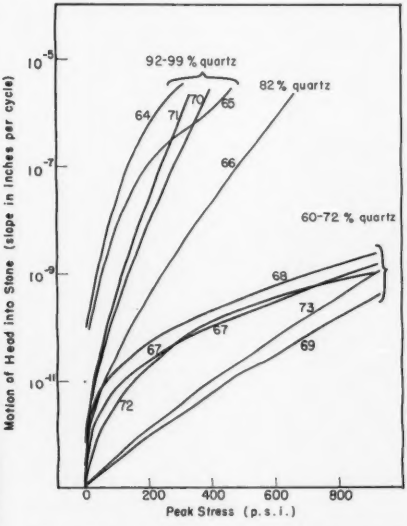

FIG. 12.—Rate of motion of loading head into crushed argillaceous and high-silica sandstones at different stress levels in the repeated load test (dry state).

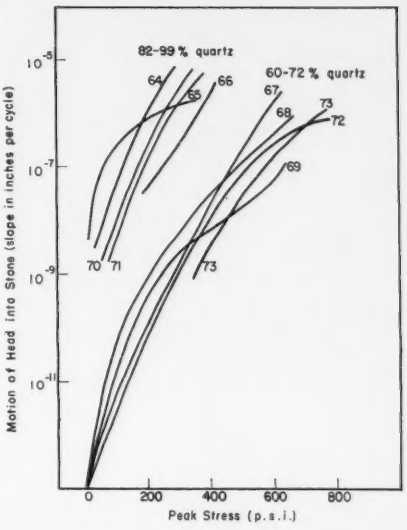

FIG. 13.—Rate of motion of loading head into crushed argillaceous and high-silica sandstone at different stress levels in the repeated load test (wet state).

rigid and brittle than the argillaceous sandstones. This is obviously a controlling factor in the fatigue test on dry stone. Since the compressive strength of solid high-silica sandstones is greater than that of argillaceous sandstones (Wolkodoff, 1955, p. 1683), the effect of rigidity of the high-silica sandstones is probably the reason for the inability of the aggregates of a crushed sample to compact or "key in." The shape of the crushed particles is also related to the packing of the aggregates. Crushed particles of high-silica sandstones tend to be more equidimensional; whereas those of argillaceous sandstones assume a flat or elongate shape because of the preferred orientation of the argillaceous material. The flat and elongate argillaceous particles, being initially poorly packed, show rapid compaction as initial cycles per load are applied; however, subsequent penetration of the head into the cylinder is slow.

The slope of degradation of high-silica sandstone is constant, and compaction of the particles is not attained at the loads applied. This may be because of the brittle

edges of the fragments. The fragments of argillaceous sandstones, being softer, abrade and compact whereas the brittle edges of high-silica fragments chip and produce more sharp edges. The fines, both those produced in the initial crushing and those produced by repeated loading, apparently behave in a manner similar to the larger fragments.

The failure of wet argillaceous sandstones may be analogous to that produced when loads are applied to sandy loam. Vomicil and others, (1957, p. 179) suggest that in dry sandy loam shera is necessary to slide the soil particles into positions where they can drop into voids, thus compacting the soil. However, in the wet soil the aggregates themselves are deformable, and the whole system is better lubricated, so pressure above causes considerable compaction.

The Proposed British Test

A static test for crushed stone, proposed in Britain, was also employed on samples 64 to 73. In this test 7 pounds of crushed stone passing the $\frac{1}{2}$-in screen and retained on the $\frac{3}{8}$-in screen are placed in a 6-in cylinder. A compressive load is then applied until the stone is compressed about $\frac{3}{4}$ in. The stone is removed, screened, and the fines passing the #8 screen are measured. The test is repeated using a new sample applying a load sufficient to produce 10 percent fine material passing the #8 screen.

This test was completed on the sandstones in both the wet and dry state. The results are shown in figure 14. The load in thousands of pounds is shown as the vertical component, and the percent interstitial material is shown as the horizontal component. The samples are plotted with increasing interstitial material from the left to right.

No direct petrographic correlation is evident since both high-silica and argillaceous sandstones have similar results. Sample 204 requires the greatest load to attain 10 percent fine material. The relatively high mica and sericite (fine-grained mica) content may have some bearing on the strength of this rock. The micas are elastic minerals, and a high proportion of these minerals would tend to give a sandstone flexibility. Flexibility, therefore, is apparently a desirable property in the British test.

The results of the British test on wet sandstone again points out the fact that water weakens argillaceous rocks but has little effect on high-silica sandstones.

CONCLUSIONS

The cohesiveness between the sand grains determines the percent wear of a sandstone under the Los Angeles test. The degree of cohesiveness is dependent upon several lithologic processes. In high-silica sandstones, cementation and pressure solution are important processes which result in an interlocking fabric and good cohesiveness between grains. In argillaceous sandstones, compaction of interstitial material determines the degree of bonding. In the calcareous and ferruginous sandstones studied, the degree of cementation is important in determining the degree of grain coherence. The degree of cementation, compaction, and pressure solution is reflected by the quantity of large interconnecting pore space.

Freezing and thawing cause a rapid breakdown of argillaceous sandstone but has little

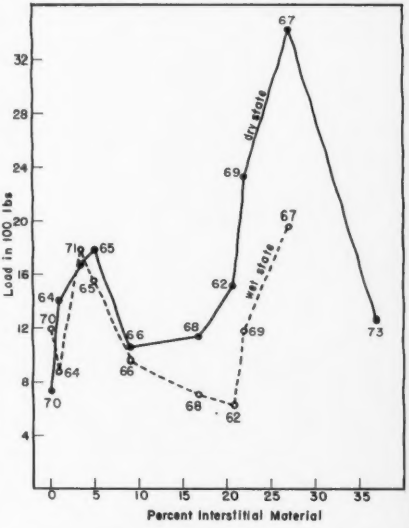

FIG. 14.—Relationship between the maximum load to give 10% fines in the proposed British test and the percent interstitial material of argillaceous and high-silica sandstones.

effect on high-silica sandstones. The larger pore space in the high-silica sandstone allows the freezing water to escape without exerting a breaking stress on the quartz grains. The type of argillaceous material is also important in determining resistance to freezing and thawing.

Compressive strength tests on argillaceous sandstones indicate that sandstones with a high percent wear also have a low compressive strength. The compressive strength of solid stone is, however, greater than that required for traffic loads.

Repeated loads of dry crushed sandstone result in a more rapid degradation of high-silica sandstone than argillaceous sandstone. Water causes a weakening of argillaceous sandstone under this fatigue test but has little effect on high-silica sandstone.

Dry crushed sandstone of high mica content requires relatively high static loads to produce 10 percent fine material in the proposed British test. Other argillaceous sandstones and the high-silica sandstones are of about equal strength in this test. The deterrent effect of water on crushed argillaceous sandstone is also shown under this static load test.

The results of the empirical tests show that water and ice have a damaging effect on argillaceous sandstones. It would seem logical that, if water comes in contact with base or subbase material, the combined effect of repeated traffic loads and freezing and thawing would produce more failure with argillaceous sandstone than with high-silica sandstone. If weathering of base and subbase material is an important factor, high-silica sandstone would be even more suitable, since quartz is more stable than clay minerals under normal weathering. The extent of subbase weathering on sandstone should be the next phase of investigation in determining the desirability of utilization of the various types of sandstone.

ACKNOWLEDGMENTS

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FIGS. 1–10, SEPTEMBER, 1961

MICRO-CROSS-LAMINATION IN UPPER KEWEENAWAN SEDIMENTS OF NORTHERN MICHIGAN[1]

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ABSTRACT

Micro-cross-lamination is the dominant type of stratification in the fine-grained sandstone and siltstone units in the Keweenawan sediments of northern Michigan. Except for its small size it is identical to "trough" or "festoon" cross-bedding and is therefore significantly different from ordinary ripple lamination. The average set of micro-cross-lamination is 2.5 inches wide, 0.5 inch thick and 8 inches long. This structure is also extremely abundant in other fine-grained sediments although it is often partly obscured by weathering or diagenesis and may not appear as typical cross-stratification. Its presence in apparently homogeneous sediments have been revealed by staining the iron or calcite cement. It is therefore quite likely that micro-cross-lamination represents a fundamental mechanism often associated with the deposition of fine grained sediments.

Associated sedimentary structures clearly indicate that this micro-cross-lamination originates in a shallow water environment repeatedly subjected to subaerial erosion. Observations of similar structures in modern streams and flood plains suggest that it develops from migrating cusp ripples.

Micro-cross-lamination constitutes an excellent criterion for paleocurrent direction. Its value is greatly enhanced by its abundance and small size since several complete sets can easily be collected in an oriented hand specimen, sawed in numerous sections and analyzed in the laboratory.

INTRODUCTION

One of the most striking features of the Upper Keweenawan sediments of northern Michigan is the abundance of extremely small-scale cross-lamination in the siltstone and fine-grained sandstone units. It constitutes the dominant type of stratification in the siltstone units of the Freda Formation which have an aggregate thickness of over 6000 feet and is abundant in the fine-grained sandstone units of the underlying Nonesuch Shale and Copper Harbor Conglomerate. When seen in vertical section this structure is similar to ripple lamination described by McKee (1939). There are, however, significant differences between the two types. According to McKee (1939, p. 75) ripple cross-lamination is "developed through deposition alone . . ." and when seen in plan view the laminae "are comparable to the contours on a topographic map—they may converge until they meet but will never cross one another." The mode of origin of the small-scale cross-lamination in the Upper Keweenawan sediments and its relation to ripple marks are not readily apparent. The strata are cusp-shape and each stratum is truncated by the succeedingly younger strata so that it is clear that both deposition and nearly contemporaneous partial erosion are involved in its formation. Except for its extremely small size it is identical to the trough or festoon cross-bedding described by Knight (1929) which is so common in many fluvial, eolian, and marine sandstones. This structure is therefore referred to as "micro-cross-lamination" in this paper in order to distinguish it from larger small-scale cross-stratification with which there is no apparent gradation nor genetic relationship. The term "rib-and-furrow" used by Stokes (1953) probably refers to the same structure but unfortunately Stokes did not recognize it as a species of cross-stratification.

[1] Manuscript received June 28, 1960.

MODE OF OCCURRENCE

Details of the occurrence and geometry of micro-cross-lamination are remarkably clear in most of the exposures of the Upper Keweenawan sediments along the southern shore of Lake Superior. Wave action has produced essentially a polished surface over broad areas of the Keweenawan rocks and excellent exposures of both vertical and plan sections of cross-stratification are numerous. In addition the size and shape of each stratum are remarkably clear and dis-

tinct because heavy minerals which occur in amounts up to 21 percent are concentrated along selected inclined strata and produce sharp color changes. Even where the sandstone is relatively homogeneous the basic red color of the Keweenawan sediments is often leached along bedding planes and accentuates details of the micro-cross-lamination that would otherwise remain obscure.

Micro-cross-lamination in the Keweenawan sediments is found only in the siltstones and fine-grained sandstones. It is most abundant in the Freda Formation (generally considered uppermost Keweenawan) which consists of over 14,000 feet of fine-grained sandstone, siltstone and shale. In the underlying Nonesuch Shale micro-cross-lamination is common in the fine-grained sandstone units and it is locally abundant in the siltstone lenses that occur in the Copper Harbor Conglomerate. The average grain size and degree of sorting of the sediments in which micro-cross-lamination occurs are remarkably persistent and show little variation, not only in all of the Upper Keweenawan sequence, but also in a number of other formations in which the structure has been observed by the writer. The average grain size is invariably between $\frac{1}{16}$ and $\frac{1}{8}$ mm. Specimens examined by mechanical analysis all fall within the "well sorted" grade of the Trask classification (Trask, 1932) and the coefficients of sorting range between 1.04 to 1.17. Figure 1 shows grain-size distribution in typical samples containing micro-cross-laminae ranging in age from Keweenawan to Recent.

The striking similarity of the average grain size and sorting coefficients in all specimens in which micro-cross-laminae occur strongly suggests a restricted set of energy conditions in which the structure may form.

In the Freda Formation micro-cross-lamination occurs in cosets[2] ranging from a few inches to 5 feet in thickness interbedded

[2] Terminology used in describing cross-stratification in this paper is adapted with minor modifications from McKee and Wier (1953). A group of essentially conformable strata or cross-strata are referred to as a set. The term "coset" is used for two or more sets of strata or cross-strata separated by original flat surfaces of erosion, non-deposition or abrupt changes in character.

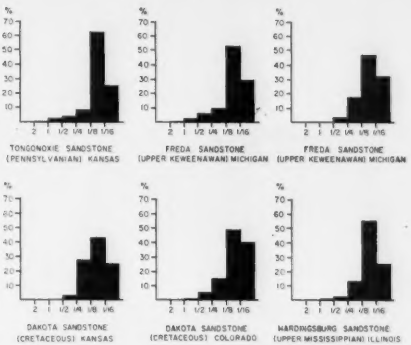


FIG. 1.—Grain-size distribution in typical samples containing micro-cross-lamination.

with cosets of horizontal laminated shale of comparable thickness (fig. 2). This simple repetition of sedimentary types persists monotonously throughout most of the 14,000 feet of the Freda Formation and represents an interesting variety of sedimentation. The aggregate thickness of the cosets of micro-cross-lamination in the Freda Formation is in the order of magnitude of 6000 feet.

The cosets of micro-cross-strata are separated from the underlying units by an erosional surface characterized by small channels which seldom exceed $\frac{1}{2}$ inch in depth. The channels invariably truncate the horizontal laminae below but fail to produce significant relief. When viewed in a vertical section the contact appears as a wavy line that can be followed laterally for many hundreds of feet. The upper surface of the cosets is characteristically a smooth planar surface parallel to the horizontal laminae of the overlying unit. Irregularities in the upper surface produced by ripple marks, however, are not uncommon.

The exact shape and size of the cosets of micro-cross-strata could not be definitely determined because no single outcrop completely exposes the complete coset in three dimensions. Most cosets can be traced laterally several hundred yards without showing significant changes in thickness or stratification. A few, however, show that the micro-cross-laminae may pass laterally into horizontal laminae. Available data

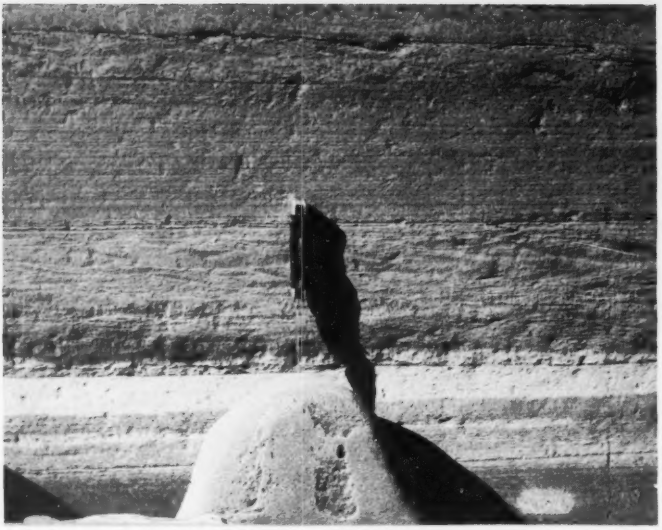


FIG. 2.—Interbedded cosets of horizontal laminae and microcross-laminae in the Freda Sandstone of Michigan.

indicate that the structure did not orginate in channels but was formed on broad flat surfaces.

DETAILS OF SHAPE AND DIMENSIONS

The small size of micro-cross-lamination permits detailed study of size and shape that is generally not easy to obtain for a larger structure. An entire coset may be brought into the laboratory and sawed into various sections exposing the internal structure and relations between individual sets. It is clear that the set of micro-cross-strata is the most basic and significant unit. The sets are shaped like a quadrant of an elongate ellipsoid and range from 1 to 6 inches in width, 0.1 to 0.8 inches in thickness and 1 to 15 inches in length (fig. 3). Over 80 percent of the sets, however, are approximately 2.5 inches wide, 0.5 inches thick, and 8 inches long. The uniform size of the structure is thus one of its most outstanding characteristics and indicates a fundamental type of mechanism associated with deposition of fine sand and silt. Significant variations in size occur mainly in the length of the structure and undoubtedly result from erosion of the older sets prior to deposition of the younger (fig. 4). It is highly probable that the original length of the sets was many times greater than what is now preserved.

Seen in vertical sections the shape of the sets of micro-cross-strata appears to be extremely variable and complex suggesting a number of different structural types. Most of the variations reflect only the apparent shape and result from different orientations of the vertical section with respect to the long axis of the structure. A classification of micro-cross-lamination as lenticular, tabular, or wedge-shape would have no significance because all three types may appear in a single set of cross-strata. In vertical sections normal to the long axis the sets appear to be lenticular with the lower surface being concave upward (fig. 5). The curvature of the lower surface is generally symmetrical but asymmetrical forms are not uncommon. The upper surface is commonly truncated by two or more sets and is characteristically scalloped. In vertical sections parallel to the long axis of the structure, the sets of micro-cross-strata appear to be tabular in shape and the trace of the lower surface is generally a straight line (fig. 5). Vertical sections

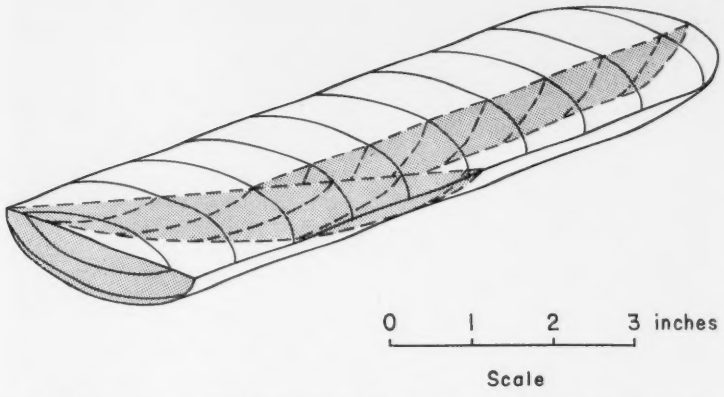


FIG. 3.—Diagram of a typical set of micro-cross-strata showing the apparent shape of the strata in plan view and in vertical sections normal, parallel, and diagonally to the long axis of the set.

FIG. 4.—Plan view showing the size and shape of micro-cross-laminae in the Freda Sandstone. Knife points in the direction of current flow.

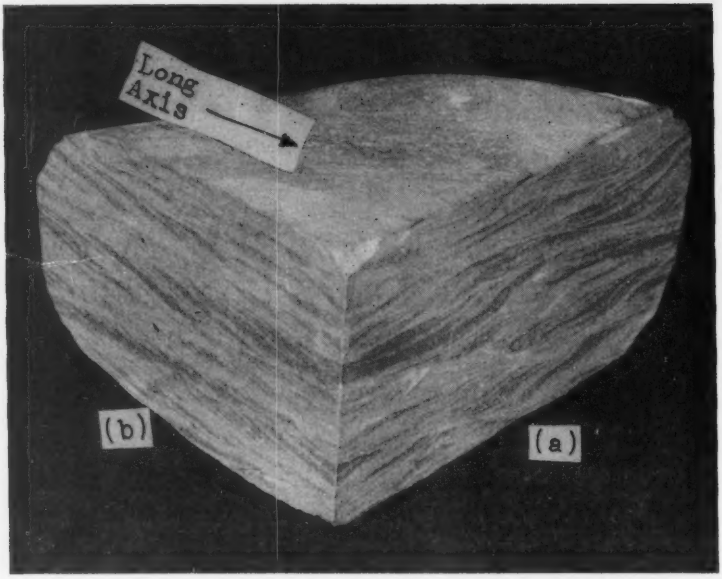


FIG. 5.—Sawed specimen of micro-cross-lamination showing the apparent shape of the sets of cross-strata in (a) a vertical section normal to the long axis of the sets and (b) in a vertical section parallel to the long axis of the sets. Note the difference in the apparent shape of the cross-strata when viewed in different vertical sections (actual size).

oriented diagonally across the axis of the structure are characteristically wedge-shaped.

The shape of the cross strata within the sets is identical to the larger trough cross-strata described by Knight (1929). In plan view the trace of the strata forms a smooth arc which is concave in a down current direction (fig. 4). The radius of curvature of the arc in most of the strata ranges from 1.5 to 2 inches. In some exposures most of the curvature occurs near the ends of the arc so that a large segment of the stratum is essentially straight or only slightly curved. Most strata are symmetrical but asymmetrical forms may result from unequal current velocities at the time of deposition or the symmetry may be destroyed by erosion that may occur prior to deposition of younger sets (fig. 6). In some exposures a single stratum can be traced across several sets forming a sinuous pattern instead of a simple arc.

Vertical sections parallel to the long axis of the structure show the micro-cross-strata dipping in a down-current direction at an average angle of 20 degrees. The angle of dip is greatest near the upper surface of the sets but it decreases rapidly in a down-current direction and almost becomes tangential with the lower surface of the set before being truncated by younger laminae. The strata are therefore characteristically concave upward but there appears to be a complete gradation into straight strata that form a distinct angular relation with both the upper and lower surfaces of the sets. Apparently the degree of curvature of the cross-strata in vertical section is proportional to the degree of curvature of the strata in plan view.

In vertical sections normal to the long axis of the structure the micro-cross-laminae are characteristically U-shaped and concave upward (fig. 5). The radius of curvature is normally about 2 inches but may range up to 10 inches depending upon the original size of the sets. The shape of micro-cross-

FIG. 6.—Plan view of micro-cross-strata showing variations in symmetry of the strata. The strata on the right are symmetrical and exhibit a smooth arch. In the center set most of the curvature is near the sides of the set and a large part of the strata is straight. The set on the left is highly asymmetrical (actual size).

strata exposed in vertical sections oriented diagonally across the long axis of the structure is transitional between that seen in longitudinal and transverse vertical sections. In diagonal sections the strata are concave upward and are characteristically asymmetrical.

Micro-cross-lamination is so small that it is easily masked by weathering and diagenesis so unless one is familiar with its size and shape the structure may escape detection. Exposures in which the structure is well expressed are further limited by the fact that in many formations changes in color, grain size and composition do not occur on a scale sufficiently small to accentuate the size and shape of the planes of stratification. The most typical expression of the structure is developed on the horizontal section. Differential weathering commonly etches out the inclined strata in relief so the structure appears as an imbrication of small troughs (fig. 7). Ridges commonly develop at the contact between two sets of cross-strata so that the structures often appear as parallel grooves separated by semi-continuous ridges. This expression is apparently the result of a greater degree of cementation along the sides of the sets where maximum permeability occurred immediately after the structure was formed.

ORIENTATION

The orientation of micro-cross-lamination is easily determined wherever the plan view of the structure is exposed and differential weathering has either etched out the laminae in relief or produced definite ridges parallel to the sides of each set of strata. The best exposures observed by the writer are found along the southern coast of Lake Superior where numerous wave cut terraces clearly expose the plan view of literally thousands of sets of strata (fig. 8). On any given stratigraphic horizon the azimuths of all micro-cross-strata are essentially parallel. The straightness and uniformity in the size of the structure is especially notable in good exposures where the sets exceed 1 foot in length. This clearly indicates that all of the micro-cross-strata on any given horizon are the product of a single current system mov-

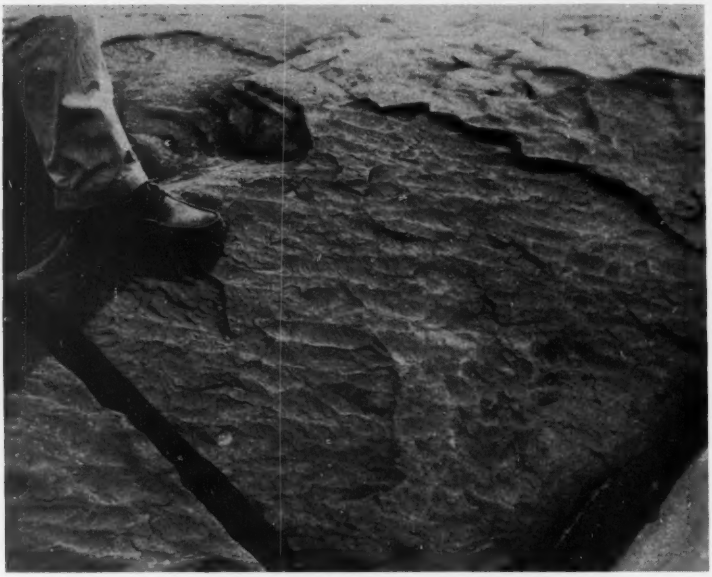


FIG. 7.—Plan view of micro-cross-strata showing typical expression of parallel ridges produced by differential cementation and differential weathering. Note the length of each set and the orientation of all the sets on one horizon. Direction of current flow was from left to right.

ing in a constant direction. The direction of paleocurrent flow as indicated by micro-cross-lamination invariably agrees with that obtained from ordinary larger-scale cross-stratification, ripple marks and current lineations (fig. 9). The absence of significant reversals ln the orientation of micro-cross-lamination throughout a section indicates that the structure was formed by currents of long duration controlled by the regional paleoslope.

ASSOCIATED SEDIMENTARY FEATURES

A number of sedimentary features are invariably associated with micro-cross-lamination and are significant in that they clearly indicate certain aspects of the depositional environment. Paramount of these are ripple marks, shale pellet conglomerate, and mud cracks.

Ripple marks.—A great variety of ripple marks are found in the Upper Keweenawan sediments and next to cross-stratification they are by far the most common and characteristic sedimentary structure. Although there is almost infinite variety in size and detail of the ripples three basic types are recognized: (1) asymmetrical ripple marks characterized by parallel ridges and troughs, (2) crescent-shaped asymmetrical ripple marks which normally appear as a series of barchan-like mounds, and (3) normal symmetrical ripple marks produced by oscillating water. Numerous modifications result from a combination of the basic types producing various eccentric patterns and interference ripple marks.

Crescent-shaped current ripple marks are particularly significant in that they are commonly found on the upper surface of the cosets of micro-cross-laminae suggesting a genetic relationship to the micro-cross-laminae. The name "cusp ripple" is used by McKee (1954, p. 64) in discussing this type of ripple mark in the Moenkopi Formation although the term "current marks" used by Kindle (1917, p. 36) and "linguoid ripple marks" used by Twenhofel (1939, p. 524) undoubtedly refer to the same structure. Others have interpreted this structure as

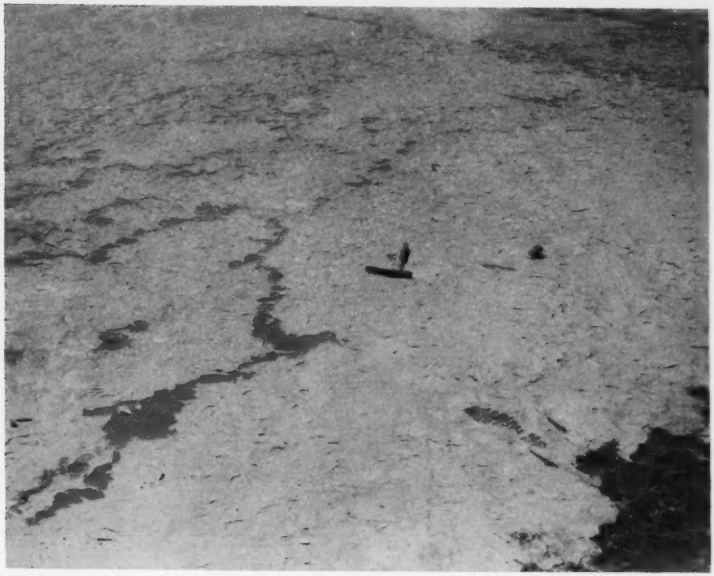
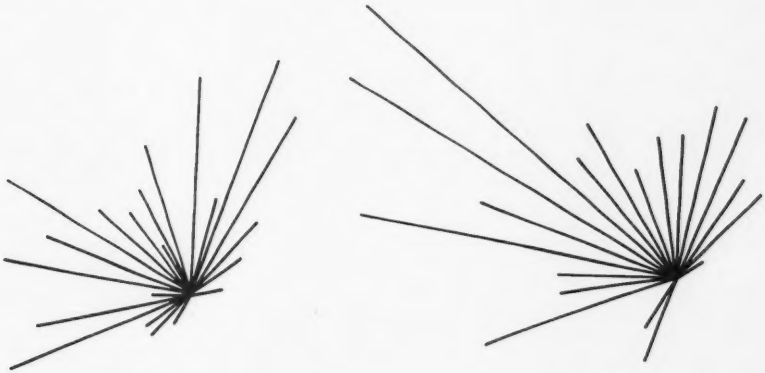


FIG. 8.—Exposure of the plan view of micro-cross-lamination in a wave-cut terrace on the west side of the Keweenaw Peninsula. Pick handle points in the directions of paleocurrent flow.



a b

FIG. 9.—a. Orientation of 217 measurements of micro-cross-lamination in the Freda sandstone.
b. Orientation of 286 measurements at medium to large-scale cross-bedding and other current structures in the Freda sandstone.

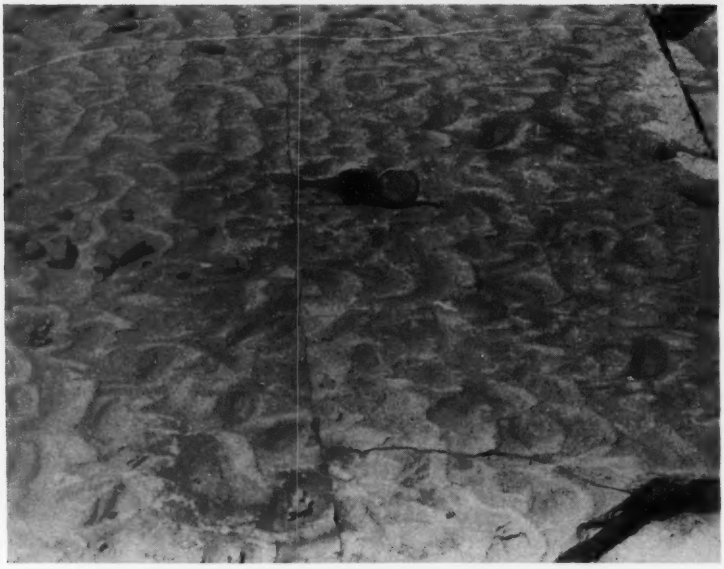


FIG. 10.—Cusp ripple marks in the Freda Sandstone of Michigan. Compass points in the directions of current flow.

interference ripple marks (Potter and Glass, 1958, p. 27).

Cusp ripple marks in the Upper Keweenawan sediments range from 1 to 6 inches in width (measured normal to the direction of current flow), .2 to 1.5 inches in amplitude and 3 to 8 inches from crest to crest (fig. 10). Although the basic shape of cusp ripple marks is similar to a barchan dune there is considerable variation in details of form and symmetry. Many cusp ripples tend to merge so that the barchan-like form is partly destroyed and the ripples appear as a series of depressions flanked by sinuous ridges. In some formations a complete gradation was observed from ripple marks with remarkably straight crests to those which form a sinuous pattern which in turn break up into crescent-shaped cusp ripples.

Shale pellet conglomerate.—Zones of shale-pellet conglomerate are common throughout the Upper Keweenawan sediments studied and are significant in that they record intraformational breaks in sedimentation in the environment where micro-cross-bedding forms. In the basal 1600 feet of the Freda Formation 145 significant zones of shale-pellet conglomerate were counted. The zones range from several inches to more than 3 feet in thickness but generally do not extend laterally more than several tens of feet. Their occurrence at hundreds of places testifies to a vast number of pauses in sedimentation accompanied by local erosion.

The individual fragments of shale range from a fraction of an inch to large blocks over 10 feet long. All of the larger particles are tabular and show little effects of transportation. Smaller pebbles, however, are normally discoidal and moderately well rounded.

In the Freda Formation the pellets are concentrated in thin lenticular zones many of which can be traced laterally into the truncated edges of a shale bed from which they were derived. Isolated "floating pebbles" of shale scattered at random throughout the section of sandstone are not uncommon.

Mud cracks.—Mud cracks are much less common in the Upper Keweenawan sediments than ripple marks or shale pellet

conglomerate but their occurrence at a number of different horizons clearly indicates that the environment in which micro-cross-lamination occurs was repeatedly subjected to subaerial conditions and periods of dessication. Mud cracks examined by the writer form a rather irregular network of polygons. The polygons range in size from 0.5 inches across to over 6 feet in diameter. Rain prints, rill marks, and current lineations are also common in the zones where mud cracks are abundant.

ORIGIN

It is apparent from the associated sedimentary structures that micro-cross-lamination in the Upper Keweenawan sediments of northern Michigan originated in a shallow water environment that was repeatedly subjected to subaerial erosion. The constant orientation of the structure indicates that the depositing current system was of long duration acting in a constant direction. The restriction of micro-cross-lamination to only the fine-grained sandstones and siltstones further indicates a limited range in the level of mechanical energy within the environment of deposition. Currents were apparently not strong enough to transport coarse sand, but were too strong to allow clay size particles to settle out. The alternation of cosets of micro-cross-laminae and cosets of horizontal laminae indicates, however, that the energy level was periodically reduced so that deposition occurred alternatingly in a standing body of water and in water moved by a weak current system. It appears that the structure originated on a flood plain or tidal flat rather than in a lacustrine or shallow sea environment.

Whether the environment was a flood plain or tidal flat or both is not readily apparent. The red color of the sediment favors the flood plain theory though it is possible that a red color could be produced under tidal waters. The constant paleocurrent direction as indicated by cross-stratification, current ripple marks and other directional sedimentary structures also favors a flood plain theory as one would expect significant reversals in current direction on a tidal flat resulting from ebb and flow of tides.

The mechanics by which micro-cross-lamination is formed is not readily apparent from the structure itself. The similarity in shape between it and larger scale trough cross-stratification suggests that they were formed by a similar mechanism; one capable of operating at vastly different energy levels. Studies of the internal structure of cusp ripple marks in the Keweenawan sediments reveals a build-up of cross-strata apparently by deposition on the lee slope as the ripple migrates downstream. The origin of cusp ripple marks, however, is poorly understood. Several early geologists failed to recognize this structure as a true ripple mark and attributed it to flows of mud (Hall, 1843; Owen, 1852). It has been observed in the process of formation on tidal flats and in small brooks by Kindle (1917) who believed that it forms "when converging currents meet or where owing to irregular contour of the bottom or sides of the channels variable or irregular currents develop." McKee (1939) shows an excellent illustration of this structure from the upper and middle portions of the Colorado River Delta and later concludes (1954) that cusp ripples develop under strong current action and result from irregular or fluctuating streams. The writer has observed cusp ripples in the process of formation along the flood plains and banks of mature streams. Micro-cross-lamination develops and is preserved if there is a constant supply of material and deposition exceeds erosion so that only the crests of the ripples are decapitated before being buried by the next ripple upstream. The basic crescent-shape of the cusp ripple is thus imparted to the inclined lamination on the lee slope so that the resulting sets of cross-strata assume a trough-like form. By this process sets of micro-cross-strata attain a length of several feet.

Flume experiments by Leopold and Wolman (1957) indicate that cusp ripple marks will form under the following conditions:

Discharge: cubic feet, per second	.033
Medium grain-size in feet	.00059
Load: pounds per second	.00121
Velocity: feet per second	.68
Slope	.00175

SIGNIFICANCE

The writer has observed micro-cross-lamination identical to that found in the

Upper Keweenawan sediments in a number of younger formations including the Hardingsburg Sandstone (Mississippian) of southern Illinois, the Pottsville Sandstone (Pennsylvanian) of Pennsylvania, the Sangre de Cristo Formation (Pennsylvanian) of Colorado, the Tonganoxie Formation (Pennsylvanian) of Kansas, the Dakota Formation (Cretaceous) of Kansas and Colorado, the Moenkopi Formation (Triassic) of Colorado Plateau, the Morrison Formation (Jurassic) of Colorado and in many recent sediments along flood plains in Kansas and Oklahoma. In all of the above formations the size and shape of micro-cross-lamination are remarkably constant and there is little variation in grain size and degree of sorting in the sediments in which it occurs. It appears therefore that micro-cross-lamination is a very common sedimentary structure in fine-grained sediments and represents a fundamental mechanism associated with deposition of siltstones and fine-grained sandstones.

The basic shape of micro-cross-lamination is identical to larger scale trough cross-stratification although it may be as much as 5000 times as small. It is therefore quite possible that trough cross-stratification, regardless of its size, is formed by one process; that is by deposition on the lee slope of migrating barchan—like sand bodies. Such features would include cusp ripples, crescent-shaped sand waves, cuspate bars, and barchan sand dunes; all of which are common in areas of recent sedimentation. The size of the structure would be governed mainly by the energy level in the environment of deposition.

Micro-cross-lamination is easily masked by weathering and diagenesis, and unless one is familiar with its size, shape and typical expression it may go undetected. In many sandstones it may not be expressed at all because changes in grain size, color, or composition do not occur on a scale sufficiently small to accentuate the form and size of the planes of stratification. In several specimens of apparently homogeneous sandstone the presence of micro-cross-lamination was revealed by staining the iron or calcite cement which was concentrated along the bedding planes.

Micro-cross-lamination appears to be completely diagnostic of water action. Air currents can produce large-scale cross-bedding, ripple marks, and lineations that have their counterparts formed by water, but it seems highly improbable that micro-cross-lamination could be produced in dry sediment by wind action. It probably can form in a number of environments (flood plain, delta, tidal flat, etc.) but it always indicates a low level of mechanical energy. In the Upper Keweenawan sediments and most other formations, rain prints and mud cracks indicate very shallow water. It is conceivable, however, that micro-cross-lamination could develop in deeper water where current action is able to move fine sediments.

Micro-cross-lamination constitutes an excellent criterion for paleocurrent direction. Its value is greatly enhanced by its abundance and small size. Whereas only a component direction can be determined from many exposures of larger-scale cross-stratification, the true direction can easily be obtained from micro-cross-lamination since an entire coset can be collected in an oriented specimen, sawed in any number of sections and analyzed in the laboratory. Micro-cross-lamination also holds great promise for studying paleocurrent directions from oriented core samples.

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FIGS. 1–13, SEPTEMBER, 1961

PETROGRAPHY OF NORTHEASTERN LAKE MICHIGAN BOTTOM SEDIMENTS[1]

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ABSTRACT

A direct relationship between sediment properties and their environment of deposition, source material, and post-glacial history was shown by textural and mineralogical studies of cores and dredge samples obtained from northeastern Lake Michigan.

The lake bottom topography of the area, which may be characterized as a valley-and-ridge province, exerts a very strong control on the distribution of sediments. In general the coarse sediments are found on the ridges and their adjacent flanks and fine sediments are restricted to the hollows between ridges. There is a regular variation in size with depth, but no regular variation in size with relation to distance from shore. The majority of bottom sediments are well sorted, but most of the samples from the lower portion of the flanks of troughs show poor sorting, probably the result of slumpage of coarser sediments from the higher portions on the flanks.

Evidence for the Lake Chippewa low-water stage is shown by the following: Core samples taken from depths of water less than 275 feet contain coarse sediment overlying, underlying, or included within a sequence of much finer grained silty clay or clayey silt. The presence of shells of shallow-water forms in some of these coarse sediment zones and a peat deposit found in association with this coarse sediment give additional evidence of a low-water condition.

The similarity of the mineralogy of the detrital sand-sized material in the lake sediments to that of the glacial drift from the adjacent shore shows that the lake sediments were derived mainly from that source.

The general order of abundance of clay minerals in both the lake sediments and glacial till material from the adjacent shore is illite, mixed-layer chlorite and illite materials, and chlorite. Little variation in the clay mineral composition with changes in depth of water, distance from shore, or depth of burial was observed.

In general the carbonate content of the lake bottom samples increases with a decrease in the median diameter of the sediment. There is, however, a lower carbonate content in the dark-gray sediments found in the deep troughs and in Little Traverse Bay: this is attributed to the solution of calcium carbonate as a result of a decrease in the pH within the sediment.

Pyrite, in association with calcareous shell material, was found in many core samples, but not in glacial till samples from the adjacent land areas. The pyrite is interpreted as secondary, resulting from local reducing conditions produced by the decay of soft parts of the shelled organisms.

INTRODUCTION

Conditions of Study

This study is a part of a program of limnological investigation of the Great Lakes sponsored by the Great Lakes Research Institute of the University of Michigan. It was carried out under the direction of Professor Jack L. Hough of the University of Illinois, a Research Associate of the Great Lakes Research Institute. The writer held an assistantship, granted by the University of Illinois Research Board, during the summer of 1958 when the field work on which this study is based was carried out.

Location

The area studied, as shown in figure 1, is located in northeastern Lake Michigan. It has been characterized as a "ridge and valley province" because it includes a number of lake-bottom valleys with depths in excess of 350 feet, separated by ridges with depths generally of 50 to 100 feet (fig. 5).

Object of Study

The objectives of this study were: (1) to determine the textural and mineralogical properties of lake sediments and of glacial tills. (2) to examine the relationship between properties of the sediment and their

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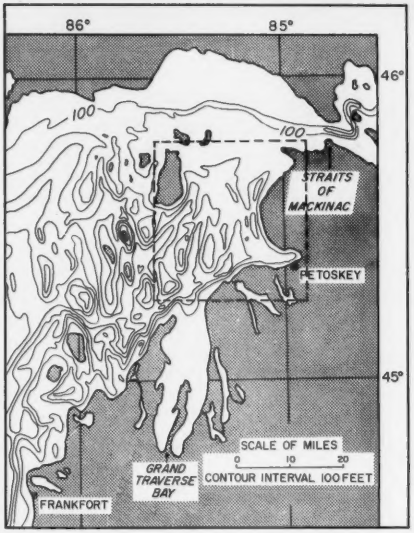

FIG. 1.—Northeastern Lake Michigan, with the area of study outlined by a dashed line.

environment of deposition. (3) to investigate the possibility that the sediment might give some indications of the details of the post-glacial history of the area. (4) to compare the sediments and glacial drift of the area because it has been assumed that the unconsolidated drift is the principal immediate source material of the sediments.

Previous Investigations

Hough (1935) in a study of southern Lake Michigan bottom deposits made the first detailed report of the recent sediments in Lake Michigan. This study initiated the modern quantitative investigation of sediments in Lake Michigan.

A number of core and dredge samples were taken from central and south central Lake Michigan during the summers of 1950 and 1951, as a part of a University of Illinois Office of Naval Research project (Hough, 1953). The mechanical composition of 67 of these core samples was reported by Snodgrass (1952). The relationship between fathogram traces of the bottom and lithology of sediments in the core samples was described by Hough (1952). An interpretation of the lithology of the cores as indicating a low stage of Lake Michigan, 350 feet below present lake surface elevation, was published by Hough in 1955.

Lauff (1957), who reported on the physical limnology of Grand Traverse Bay, made mechanical analyses of 27 samples and correlated the activity of currents in the bay with the distribution and character of the bottom sediments.

Hough (1958) presented a summary of the physical environment, the bottom sediments, and the geologic history of Lake Michigan, as part of a treatise on the geology of the Great Lakes.

Rogers (1959) studied four core samples of deep water sediments from northwestern Lake Huron, reporting the texture, carbonate content, and clay mineral composition of these sediments.

The U. S. Lake Survey publishes annually the Great Lakes Pilot, which gives detailed descriptions of various features pertaining to navigation, and it has published a number of navigation charts which show water depths in considerable detail, and include the area of the present study (Corps of Engineers, U. S. Army, 1959).

Maps of the glacial deposits in the Southern Peninsula of Michigan have been prepared by Leverett and Taylor (1915) and Martin (1955). A map showing the extent of Valders glaciation in Michigan was prepared by Melhorn in 1956.

Geologic Setting

The Michigan structural basin, which is nearly circular in outline, is centered in the lower peninsula of Michigan. Lakes Michigan and Huron are located concentrically around the flanks of this structural basin. They lie, for the most part, in strike valleys of shales and other relatively weak rocks. The bedrock of the area studied, and immediately adjacent areas includes the Paleozoic sediments listed in table 1. This area lies on the northwest flank of the Michigan structural basin, and the bedrock formations have an east-west regional strike and a southward dip.

The Geologic Map of the Southern Peninsula of Michigan (Martin, 1936) indicates that the entire area of study lies within the outcrop belt of the Lower and Middle

TABLE 1.—*Classification of bedrock formations in the northwestern portion of the Michigan structural basin*[3]

System	Group	Formation	
DEVONIAN	Traverse	Petoskey Limestone Bell shale	
		Rogers City Limestone	
	Cazanovia	Dundee Limestone	
	Detroit River	(undifferentiated)	Mackinac Breccia
	Onesquethaw	Bois Blanc	
	Deerpark	Garden Island	
SILURIAN	Bass Island	St. Ignace Dolomite	
	Salina	Point aux Chenes Shale	

[3] Ehlers, Smith, and Shelden, 1959, p. 12.

Devonian formations. The shoreline from Charlevoix to Bay View in Little Traverse Bay is formed by a cuesta of the Petoskey Limestone of the Traverse Group (Middle Devonian).

The area of the present study lies in the ridge and valley province of Lake Michigan (Hough, 1958, p. 25). The bottom topography in this area consists of numerous deep valleys and ridges which do not coincide with the regional strike of the bedrock formations. Most of these valleys have a north-south trend. According to Landes, Ehlers, and Stanley (1945, pp. 143–145) this unusual structural phenomenon has been produced by ground water leaching large quantities of salt from the Salina Formation around the margin of the Michigan structural basin. The rocks overlying the salt collapsed or settled downward producing large fault blocks.

After the Paleozoic Era and prior to the Pleistocene Epoch, the Michigan basin was probably a broad stream valley located in the outcrop belts of the weaker rock formations of the area. This preglacial lowland exerted a strong control over the direction of ice movement and it seems likely that the glacial ice moved essentially parallel to the trough axes in the area of study.

PLEISTOCENE AND RECENT HISTORY OF NORTHERN LAKE MICHIGAN

General

The Pleistocene to Recent history of the region has been described by Hough (1958) and the following summary of the parts of that history which are pertinent to the present study has been taken mainly from that publication.

Continental glacial ice covered the area of study during each of the four major glacial ages of the Pleistocene Epoch, and the ice also covered the area during each of the subages of the latest, or Wisconsin Age. Because each advance of ice over the area tended to destroy or at least to obscure any evidence of former glacial and interglacial events within the area, only the later events need be described here.

Late Pleistocene Events

The classification of the Wisconsin subage which is used in the present study is based on a generally-accepted classification developed by Leighton (1933, p. 168, and 1957, p. 108–109) which has been adapted for use in the Great Lakes region by Hough (1958, p. 94). Table 2 shows the late Wisconsin time scale for the area of study. Leighton's "Mankato Subage" has been replaced by "Port Huron Subage"; the two names probably are correlative but the name based on the local Port Huron morainic belt is considered preferable in order to avoid confusion. The Port Huron is considered as a subage separate from the Cary because the Port Huron morainic system in Michigan marks a pronounced readvance of the ice front and it was described by Leverett and Taylor (1915, p. 293) as "of the substage order of magnitude, and is, in fact, one of the best developed and most

TABLE 2.—*Late Wisconsin time scale for the area of study*[4]

Recent ("post-glacial" of this area)
Valders Glacial Subage
(Two Creeks Interval)
Port Huron Glacial Subage
(Cary-Port Huron Interval)
Cary Glacial Subage

[4] Adapted from Hough, 1958, p. 94.

clearly defined moraines in the Great Lakes region."

The Cary-Port Huron Interval was a time of marked retreat of the ice-front in the Great Lakes region, as shown by the fact that the Port Huron moraine records a pronounced readvance, and by evidence for a low-water lake stage in the Erie basin at this time (G. W. White, as quoted by Hough, 1958, p. 147). The extent of ice-front retreat during the Cary-Port Huron interval in the area of study is not known, but Hough (1958, p. 168–169) has suggested that it was sufficient to allow discharge from the northeastern part of the Michigan basin to the Huron Basin.

The Two Creeks Interval is well established by the section including the Two Creeks forest bed, north of Manitowoc, Wisconsin (Wilson, 1932, 1936; Hough, 1958, p. 102–103). During this interval the ice-front retreated far enough to permit northeastward discharge from the Michigan basin and to cause a lowering of water level in the lake to a level below the present lake level. Additional evidence for an extensive retreat at this time is the color and composition of the till deposited during the subsequent Valders Glacial Substage (Murray, 1953, p. 148–154).

The Valders, originally named by Thwaites (1943, p. 121), and represented by the red till in northeastern Wisconsin, is believed also to be represented by the red drift of the northern part of the southern peninsula of Michigan (Bretz, 1951, p. 423–424; Melhorn, 1954; Zumberge, 1956; Hough, 1958, p. 103–108; Moore, 1958). This red drift represents an extensive advance of glacial ice in the Michigan basin. The ice advanced over the area of the present study and in general to the inner margin of the Port Huron morainic belt in adjacent parts of Michigan.

The post-glacial history of the area of study began with the retreat of the margin of the Valders ice across the area.

Post-Glacial Lake Stages

The history of known lake stages in the Michigan basin began during the Cary Glacial Subage when the ice front retreated northward into the basin and water was ponded between the ice and the divide at the south end of the basin. The details of the earlier lake stages are of no concern to the present study.

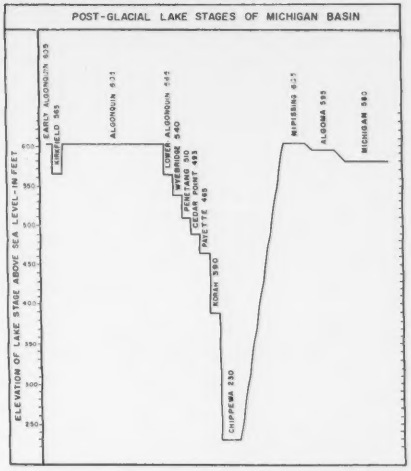

FIG. 2.—Summary of post-glacial lake stages in northeastern Lake Michigan (adapted from Hough, 1958, p. 166).

By sometime after the maximum of the Valders Glacial Subage, when the ice margin was retreating in the area of study, water level in the Michigan basin had been lowered to the Algonquin level, 605 feet above sea level. Figure 2 shows a diagrammatic summary of lake stages in the area, from the Algonquin to the present, as deduced from evidences in various parts of the Great Lakes region. The major changes in water surface elevation in the area resulted from: (1) discharge through new, lower outlets to the east and northeast when retreat of the ice margin in Ontario uncovered the dischargeways. (2) possible, on one occasion, damming of an eastern outlet by readvance of the ice front. (3) uplift of an outlet channel when the region to the northeast was raised, as a result of elastic rebound as the weight of the glacial ice was removed. (4) downcutting of the Port Huron outlet, at the south end of the Lake Huron basin.

In addition to changes in elevation of the lake surface, the entire area of study has undergone uplift during the Algonquin to present period. The actual depths of water

and positions of shorelines in the area were determined by both of these factors. Because these details of lake history may have had an important effect on the bottom materials which may be found in the area today, they are described in the following paragraphs.

Early Lake Algonquin.—The first post-Valders-maximum lake stage of the area probably came into existence while the lake surface in the Michigan basin stood at 605 feet above sea level, and the retreat of the ice front simply allowed the lake water to spread over the area. This first lake stage, at 605 feet, probably was the Early Algonquin (Hough, 1958, p. 183), the equivalent of the Toleston (Leverett and Taylor, 1915, p. 356).

The Kirkfield Stage.—A Kirkfield stage, during which lake level was 40 to 60 feet lower than the Algonquin, has been deduced mainly from evidence in the southern Georgian Bay area of Lake Huron and in the vicinity of Kirkfield, Ontario (Leverett and Taylor, 1915, p. 409–413; Johnston, 1916, p. 15–17; Stanley, 1938a, p. 483–490). This lake stage discharged through an outlet near Kirkfield that was uncovered by retreat of the ice-front in that vicinity.

Lake Algonquin.—Following the Kirkfield low stage, water level in the Huron and Michigan basins rose back to the 605-foot elevation and initiated the main Algonquin stage. The cause of the rise was the damming of the Kirkfield outlet by a local readvance of the ice, according to Johnston (1916) and Hough (1958, p. 213–214). Whatever the cause may have been, there is ample evidence for the rise (*see*, in particular, Stanley, 1938a).

Lake Chippewa.—Northward retreat of the glacial ice margin in the region east of Georgian Bay (east of Lake Huron) uncovered a series of progressively lower outlets. As a result of this, lake level in the Huron and Michigan basins was alternately lowered and held for a time at each of several progressively lower stages. In the Huron basin, the various stages of this series have been recorded by shore features and they include the stages listed in table 3.

The stages listed in table 3 are based mainly on work by Stanley (1936, 1937) and in part on work by Hough (1958, p. 225–236). In the Michigan basin very few of the stages in this series have been recognized, but the lowest level attained by the lake has been determined from evidence found in lake bottom core samples (Hough, 1955).

TABLE 3.—*Lake stages in the Huron basin (Post-Algonquin to Stanley)*[5]

Lake stage	Estimated Original Altitude above sea level (feet)
Upper Group	605–555
Wyebridge	540
Penetang	510
Cedar Point	493
Payette	465
Sheguiandah	?
Korah	?
Stanley	185

[5] Adapted from Hough, 1958, p. 227.

The lowest stage in the Michigan basin, named the Chippewa, included a lake in the main northern basin with a surface elevation of 230 feet above sea level; 350 feet below present lake level). This lake drained to Lake Stanley in the Huron basin. A map showing the approximate configuration of Lake Chippewa (Hough, 1955, p. 966) indicates that the lake extended into the area of the present study, in each of several of the deep valleys.

Lake Nipissing.—Post-Algonquin uplift of the land apparently was accelerated after the ice-margin retreated north of the North Bay outlet. Uplift of that outlet resulted in a rise of lake surface in the Huron and Michigan basins, which culminated in a return of the lake surface to the old Algonquin stage level of 605 feet above sea level. Then, even though the land in the north continued to rise, the lake level remained nearly static because of discharge through the old southern outlets at Port Huron and at Chicago. The resulting stage is named the Nipissing. The new Lake Nipissing margins reoccupied the old Algonquin shores in the unwarped southern portions of the lake basins, but they formed new shore features in the north (including the area of the present study), where the land had risen and raised the Algonquin shores above the original Algonquin level.

Lake Algoma.—The Algoma stage is recorded by well-developed shore features

which are intermediate in level between the Nipissing and the present shores and which are at an elevation of 596 feet above sea level in the unwarped southern portions of the lake basins. The lowering of lake level from the Nipissing to the Algoma stage apparently was a result of downcutting of the outlet. The static level of the Algoma stage may represent cessation of downcutting.

Transition to the Present Lakes.—Further downcutting of the Port Huron outlet resulted in lowering the lake level to 580 feet above sea level, to initiate the present stage of Lakes Huron and Michigan.

Uplift of the Land

The main Algonquin beach is horizontal, at an elevation of 605 feet above sea level, in the southern parts of both the Michigan and Huron basins; but it rises to the northward and reaches an altitude of 1015 feet above sea level at a point six miles north of Sault Ste. Marie, Ontario. At that locality, there has been 410 feet of uplift in post-Algonquin time. All of the area included in the present study was involved in the uplift. It will be necessary, therefore, to take the uplift into account in an evaluation of the data of this study.

It has been stated in a foregoing paragraph that the Algonquin stage, with its strongly-developed beaches, apparently represented a time in which little or no uplift was occurring. In his study of a series of post-Algonquin beaches, Stanley found that 18.7 percent of the total post-Algonquin uplift had occurred by Payette time (Hough, 1958, p. 230). The amount of uplift of the Nipissing beach in the area of study, about 10 feet at Petoskey, Michigan, indicates that 10 percent of the 100 feet of total post-Algonquin uplift at that locality has occurred since Nipissing time. It may be concluded, therefore, that a little over 70 percent of the post-Algonquin uplift occurred between Payette and Nipissing times. The rates of uplift within that period are not known. It is impossible, therefore, to calculate the exact positions at which the Chippewa shore line would have been located on the present bottom in the area of study. A rough estimate is made, however, by assuming that 20 percent of the total post-Algonquin uplift had occurred by the beginning of Chippewa time; then, along the isobase representing 100 feet of uplift to the Algonquin beach, which passes through Petoskey and through the south-central part of the area of study, an uplift of 80 feet would have raised the Chippewa shore from its original elevation of 230 feet above sea level to a present elevation of 310 feet above sea level. The present depth of the Chippewa shore (or the warped Chippewa water-level plane) therefore should be about 270 feet below present lake level, along the isobase passing through Petoskey (fig. 3).

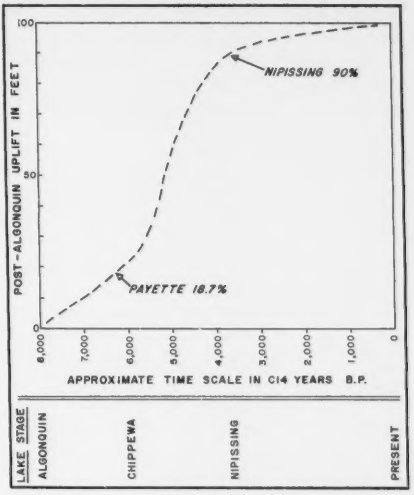

FIG. 3.—Estimated time-uplift curve for a 100-foot Algonquin isobase through Petoskey, Michigan. The two end points are known. The Payette percentage is adapted from Georgian Bay data, and the Payette date is estimated. The Nipissing percentage is derived from the Nipissing beach elevation at Petoskey, and the Nipissing date is estimated.

To further explain the combined effects of changes in altitude of the lake surface and changes in altitude of the land, the following has been done, For convenience, the effects are investigated for the line representing 100 feet of uplift of the Algonquin beach. This line passes through Petoskey and has an orientation of 110 degrees; it therefore extends through the south-central part of the area of study and crosses one of the

sampled profiles. Given the information and estimates illustrated in figures 2 and 3, the following may be stated. For any point on the present 100 foot Algonquin isobase: (1) in Algonquin time the point was 125 feet deeper (because lake level was 25 feet higher, and the land was 100 feet lower). (2) in Chippewa time the point was 270 feet shallower (because lake level was 350 feet lower, and the land was 80 feet lower). (3) in Nipissing time the point was 35 feet deeper (because lake level was 25 feet higher, and the land was 10 feet lower).

In regard to item (2) in the foregoing statement, it is obvious that those points which are at a present depth of 270 feet were at the shoreline in Chippewa time and that all points at present depths less than 270 feet were above water in Chippewa time.

PHYSICAL ENVIRONMENT

General

Lake Michigan, with a surface area of 22,400 square miles, is the third largest of the five Great Lakes. The area of drainage (excluding the lake) is 45,460 square miles. The climate of the Michigan basin alternates between continental and semi-marine depending upon the meteorological conditions. The mean annual temperature is 40° F with an average temperature in July of 67° F and an average temperature in January of 21° F (U. S. Department of Agriculture, 1941). The mean average rainfall is 31 inches. The lake water is partially frozen (especially in protected embayments) from mid- December until mid-April of the average year (Hough, 1958, p. 49). The main discharge outlet of the lake basin is at the Straits of Mackinac, where water from Lake Michigan flows into Lake Huron.

Lake Bottom Topography

The bottom topography of the southern portion of the area included in the present study (fig. 5) is somewhat unusual. The area has been called the "Ridge and Valley Province" of Lake Michigan by Hough (1958, p. 23). The complex ridge and valley topography, as shown in figure 5, consists of several deep troughs most of which have a distinct north-south trend and depths of 250 to 350 feet. These troughs are separated by ridges which are covered only 25–100 feet of water. The deep troughs of the area are connected in what appears to be a submerged drainage system, though the greatest depths are well below any known low-water level of the lake. The deepest trough extends along the south shore from a point west of Petoskey, west-southwestward along the northern face of the Petoskey Limestone cuesta, across the mouth of Grand Traverse Bay, and westward to the deep northern basin of Lake Michigan. The depth in this trough increases fairly regularly westward. Most of the troughs in this area have a generally north-south trend and they become shallower to the northward. The northern portion of the area of study has a different character. The lake topography in this area shows little variation in relief and is essentially a shelf with an average depth of 100 feet.

The shoreline from Charlevoix to Bay View is formed by a limestone cusesta of the Petoskey Limestone. This cuesta is partially submerged in water in the Charlevoix area. Beaver Island and the shoreline from Bay View north are underlain by limestone and dolomite of the Detroit River group. The position of the shoreline south of Little Traverse Bay is controlled by limestone bed rock, but north of this bay the position is more closely controlled by glacial drift material.

Lake Water

The surface currents of Lake Michigan are produced primarily by wind action. The net current flow from the basin is eastward through the Straits of Mackinac, however, this flow may be reversed due to changes in wind direction. Studies by Harrington (1895) and Ayers and others (1958) show that the surface current generally is to the northeast along the eastern side of the lake, and the surface flow generally includes a counter clockwise swirl around the Beaver Island group. The northward flow east of Beaver Island, in the area of study, may be reversed when the wind in the Mackinac Straits area blows from the north to the northeast. A small clockwise eddy has been observed in Little Traverse Bay (Ayers and others, 1958).

The waves in Lake Michigan may reach

heights as great as 15 feet and a length of over 100 feet as shown by Hough (1953). The effectiveness of wave action is shown by the exposed steep sea cliffs composed of glacial till and broad beaches along the eastern shore of the lake in the area of study.

The effectiveness of wave action on the bottom is a controversial matter and no depth can be assigned to wave base in Lake Michigan (Hough, 1958, pp. 33–35). Vigorous wave abrasion has been observed in the area of investigation. A shallow spot (35–50 feet), nine miles northwest of Charlevoix, during times of high waves has a tendency to cause incipient breaking of waves. Local navigators have observed breaking waves, foam, and mist at this spot during storms.

The temperature of the lake surface water ranges from 32° F in the winter to greater than 70° F in the summer, but the temperature of the water in the deeper portions of the lake remains near 39° F throughout the year (Hough, 1958). The temperature and density distribution of water in the Michigan basin follows a seasonal pattern. During the Spring and Fall seasons the water is isothermal (same temperature throughout). During this period the water has a temperature close to the temperature of maximum density. Because the entire column of water at these times has the same temperature and same density, any wind action on the surface of the lake tends to produce complete circulation of water. During the summer there is a density stratification of the water. During this period the top lighter water of the lake (epilimnion) is isolated from the denser bottom layer of water (hypolimnion). The circulation of water tends to be restricted to the epilimnion during this period and there is little circulation of surface water to the deeper portions of the lake. During the winter the surface water is colder than the temperature of maximum density and a distinct density stratification usually develops.

The annual cycle described in the foregoing paragraph is an approximation of conditions which have been observed in the main basin of Lake Michigan (Church, 1942, 1945; Ayers and others, 1958; Hough 1958, pp. 50–57). It is modified by local conditions of upwelling and downsinking, which generally do not reach to the greatest depths of the lake, and variations in wind stress. It may be expected that the lake water in the area of study undergoes the essential features of the typical annual cycle, but there may be important local variations caused by the shoreline configuration and bottom topography.

Studies by Stimpson (1870) and Ayers and others (1958) show that the waters of Lake Michigan are completely fresh in both deep and shallow areas. The total dissolved solids in the lake water are 118 parts per million which compares more closely with the total for rain water, 40 p.p.m., than it does with the total for sea water of 35,000 parts per million (Hough, 1958, p. 60). The major constituents of the water in the order of abundance of dissolved solids are bicarbonate, calcium, silica, magnesium, and sulphate (Ayers and others, 1958). The water is alkaline in character and the approximate pH is 8.

Lake Michigan is classed as an oligotrophic lake because the quantity and rate of production of flora and fauna in the lake is very meager.

Source Material

The shores and drainage area of northeastern Lake Michigan are composed of glacial drift deposits, Recent lake clay deposits, beach sand and gravel, limestone, dolomite, and shale. The adjacent land area (northern part of Southern Peninsula of Michigan) is in the youthful stage of the geomorphic cycle. The moraines and till plains show very little erosion except in a few isolated areas. The drainage in this area is poorly developed as shown by the numerous swamp areas, many lakes, and lack of an integrated network of stream channels.

The main source of sediment in the area of study is the unconsolidated deposits of glacial till and glacial outwash that lie near the lake shore or in shallow water in the lake. These deposits due to their loose consolidation are very susceptible to wave attack and erosion.

FIELD OPERATIONS

Methods of Sampling

Field operations were carried out in the summer of 1958 with the Great Lakes Research Institute research vessel NAIAD,

a 34-foot power boat fitted for limnological work. Navigation generally was by dead reckoning, using compass courses and speed of travel. Locations of sampling stations generally were determined by observing two sextant angles between known points on the shores. The depth of water at each station was read from the vessel's fathometer.

The types of bottom sampling apparatus used in this study were an "orange-peel dredge" which has four jaws and forms a hemisphere when closed, and a Trask coring tube. Both these sampling devices were lowered and retrieved by use of an electric powered winch. After an orange-peel sample was obtained, it was placed in a metal tray and a general megascopic description as made of the contents. A portion of the sample was then placed in a pint Mason jar for shipment to the laboratory. Immediately after a core sample was taken, the plastic core liner was removed from the core barrel and rubber corks were inserted and taped to the open ends of the plastic liner to preserve the core for delivery to the laboratory.

Location of Samples

Figure 4 shows the locations of the stations of all bottom samples. The type of sample obtained is indicated at its location by a symbol: a circle represents a dredge sample and a square represents a core sample. A total of 35 dredge samples and 21 core samples were taken, the latter ranging in length from 5 to 112 centimeters.

The till samples which were used for comparison were collected from end moraines, ground moraines, and drumlins in the northern part of the Southern Peninsula of Michigan at localities shown in figure 1 and table 2 of a report by the author (Moore, 1958).

LABORATORY STUDY

Megascopic Description of Cores

The wet cores were extruded from the plastic liners and laid in split cardboard mailing tubes soon after their arrival in the laboratory. Full scale logs were then drawn indicating the color, texture, and organic remains of the sample. These core logs served as a guide in selection of samples for laboratory analyses and for correlation purposes. After the cores had been allowed to dry thoroughly in the air, portions were selected for detailed analysis. The sample intervals were determined on the basis of change in both texture and color. The cores were also sampled every 30 centimeters, within all of the thicker zones. The sampled interval ranged in length from 2 to 10 centimeters, depending on the thickness of a textural or color zone.

Mechanical Analysis

All samples selected for mechanical analysis were partially disaggregated and approximately 40 to 50 grams were separated from the bulk sample with a Jones splitter. The hygroscopic moisture was removed by placing the sample in a drying oven at a temperature of 90° F for four hours. After the sample had cooled in a desiccator and had been weighed, it was placed in a beaker containing 125 cc of 4 percent stock solution of sodium hexametaphosphate. The sample was allowed to soak for approximately 24 hours to deflocculate and disaggregate the material. During this period, the mixture was stirred several times to insure further disaggregation. Following the 24-hour soaking, the sediment was transferred to a metal container and agitated for two minutes on a Hamilton Beach mixmaster.

The sample was then wet sieved through the .062 mm sieve by washing gently with distilled water. The material retained on this sieve was dried and sieved for 10 minutes on a Ro-tap machine in a set of Tyler screens having openings ranging from 2.00 to .062 mm. The fractions thus separated were weighed and the weights were entered on a data sheet.

The material passing the .062 mm sieve was transferred to a 1000 ml cylinder and the cylinder was filled to 1000 ml with distilled water. The grain size distribution of this material was determined with a standard soil hydrometer, using the method of Bauer and Thornburn (1958). Hydrometer readings were taken at 2, 4, 8, 15, 30, and 60 minutes and at 2, 4, 8, and 24 hours.

The calculations of the hydrometer analyses were performed by the Illiac digital computer, which greatly decreased the time required for calculation. The data obtained from the sieve and hydrometer analyses were placed on semi-logrithmic graph paper and cumulative curves were constructed

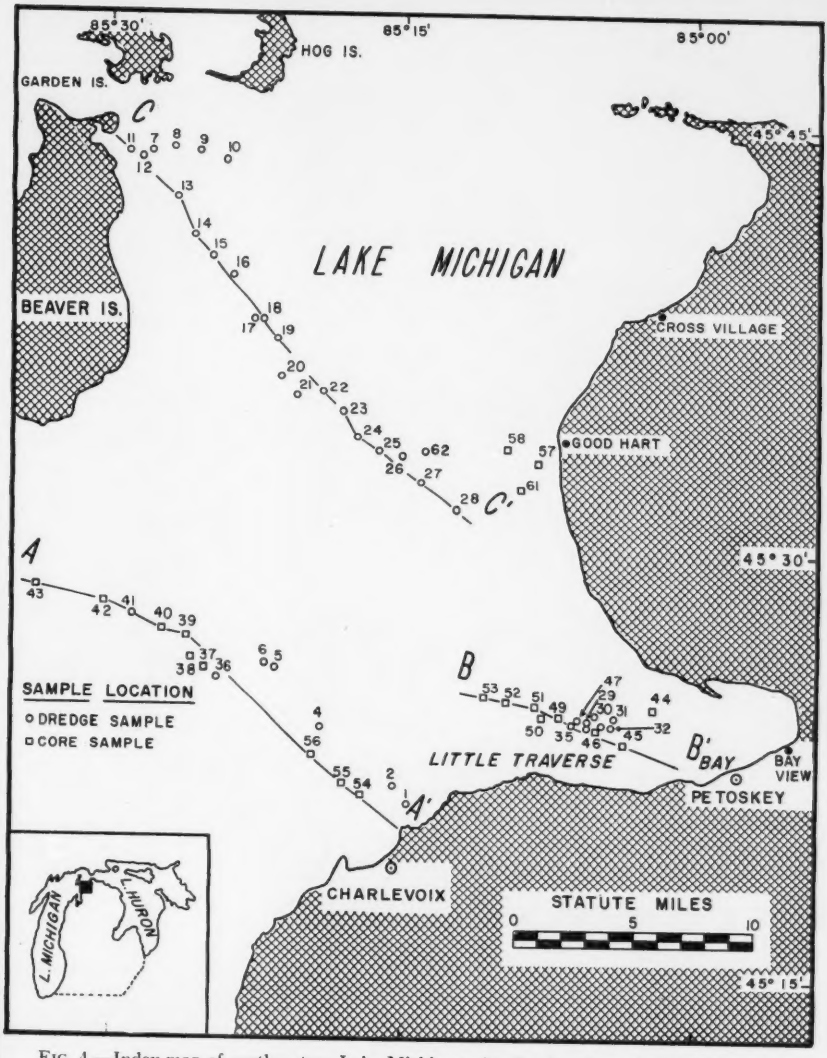

FIG. 4.—Index map of northeastern Lake Michigan, showing locations of sampled profiles.

The median, quartile values, sorting factor, and skewness were determined from these curves (table 4). Percentages of gravel (>2 mm), sand (2–.062 mm), silt (.062–.0039 mm) and clay (<.0039 mm) were also calculated (table 4). Size analyses were obtained for 141 lake bottom samples and 20 glacial till samples.

Microscopic Examination of Sand Fraction

The material in the sample greater than .062 mm was studied under the binocular

TABLE 4.—*Percentage mechanical composition and statistical constants of northeastern Lake Michigan samples*

Sample number	Gravel	Sand	Silt	Clay	Q_3 mm)	Md (mm)	Q_2 (mm)	So
M-58-1	1	86	10	3	.210	.125	.086	1.54
M-58-2	1	59	20	20	.160	.078	.008	4.4
M-58-4	1	17	32	50	.020	.004	.001	5.3
M-58-6	70	30	—		10.0	5.0	.660	3.9
M-58-7	5	81	10	4	.330	.20	.094	1.81
M-58-8	100	—	—	—	—	—	—	—
M-58-9	—	100	—	—	.380	.280	.220	1.3
M-58-10	3	94	2	1	.335	.220	.152	1.48
M-55-11	100	—	—	—	—	—	—	—
M-58-12	1	99	—	—	.330	.246	.190	1.32
M-58-13	1	99	—	—	.335	.310	.195	1.31
M-58-14	73	27	—	—	30.0	13.0	1.5	4.5
M-58-15	—	100	—	—	.325	.250	.185	1.28
M-58-16	2	83	9	6	.330	.230	.115	2.87
M-58-17	—	92	4	4	.355	.270	.205	1.31
M-58-18	45	52	3	—	—	.900	.250	—
M-58-19	100	—	—	—	—	—	—	—
M-58-20	—	48	24	28	.220	.050	.003	9.46
M-58-21	100	—	—	—	—	—	—	—
M-58-22	4	64	14	18	.280	.162	.002	3.66
M-58-23	3	67	12	18	.330	.182	.025	3.63
M-58-24	8	77	11	4	.440	.280	.135	1.8
M-58-25	1	87	6	6	.420	.300	.150	2.22
M-58-26	—	79	9	12	.320	.194	.068	1.64
M-58-27	—	52	24	24	.220	.074	.005	6.84
M-58-28	—	33	28	39	.130	.009	.002	9.25
M-58-29	1	88	9	2	.370	.260	.180	1.43
M-58-30	—	94	5	1	.350	.260	.160	1.47
M-58-31	—	42	31	27	.200	.035	.003	8.3
M-58-32	—	10	56	34	.025	.009	.002	3.35
M-58-33	—	49	30	21	.230	.060	.006	6.17
M-58-34	1	93	3	3	.400	.280	.210	1.37
M-58-35	2	88	6	4	.340	.265	.185	1.35
M-58-36	—	90	8	2	.350	.250	.132	1.67
M-58-37								
0–6	2	20	27	51	.033	.004	.001	5.5
10–15	2	47	29	22	.280	.066	.005	7.2
20–25	2	50	25	23	.270	.070	.005	7.2
30–34	17	49	19	15	.560	.190	.020	5.27
40–45	15	40	24	21	.440	.108	.007	7.92
50–54	14	51	19	16	.880	.200	.020	6.60
M-58-38								
0–8	—	9	48	43	024	.008	.002	3.3
20–30	—	2	52	46	.012	.005	.001	3.01
50–60	—	1	53	46	.010	.005	.001	2.76
75–85	—	1	50	49	.008	.004	.001	2.7
M-58-39								
0–3	—	24	56	20	.094	.032	.006	3.96
10–20	—	3	55	42	.013	.007	.002	2.94
40–50	—	1	44	55	.010	.003	.001	2.97
60–70	—	2	46	52	.009	.004	.001	3.19
80–88	—	2	32	66	.006	.002	.001	2.98
M-58-40								
0–3	—	92	3	5	.320	.238	.135	1.53
10–20	—	2	36	62	.006	.003	.001	2.26
35–40	—	6	35	59	.010	.003	.001	3.14
42–47	—	27	35	38	.068	.012	.001	8.20
48–53	—	16	36	48	.035	.005	.001	6.20
53–56	—	52	24	24	.165	.064	.005	5.90

TABLE 4—*Continued*

Sample number	Gravel	Sand	Silt	Clay	Q_3 (mm)	Md (mm)	Q_2 (mm)	So
M-58-42								
0–2	—	86	5	9	.362	.260	.132	1.67
5–10	—	5	27	68	.010	.002	.001	3.60
12–15	12	42	21	25	.440	.125	.004	10.00
M-58-43								
0–10	—	1	56	43	.014	.006	.002	3.20
20–30	—	2	56	42	.013	.005	.002	2.90
50–60	—	2	50	48	.011	.004	.002	2.50
75–87	—	1	48	51	.011	.004	.001	3.74
M-58-44								
0–10	—	4	62	34	.016	.009	.002	2.82
30–40	—	4	65	31	.018	.007	.003	2.54
50–60	—	4	64	32	.018	.008	.003	2.68
75–82	—	5	65	30	.019	.009	.003	2.50
M-58-45								
0–10	—	3	74	23	.034	.018	.004	2.90
30–40	—	4	62	34	.018	.008	.002	2.80
50–60	—	4	62	34	.017	.008	.002	2.80
75–83	—	5	59	36	.017	.007	.002	2.80
M-58-46								
0–3	—	62	27	11	.290	.130	.023	3.50
20–30	—	12	48	40	.020	.009	.002	3.60
50–60	—	6	42	52	.012	.004	.001	3.15
70–80	—	5	39	56	.009	.003	.001	2.96
95–98	1	24	40	35	.058	.012	.002	6.00
100–105	1	24	42	33	.052	.015	.002	5.40
106–109	2	49	29	20	.380	.076	.008	7.10
M-58-47	—	84	11	5	.360	.290	.140	1.60
M-58-49								
0–3	—	76	14	10	.310	.177	.056	2.34
20–30	—	8	44	48	.015	.004	.001	3.68
50–60	—	6	36	58	.010	.003	.001	2.96
61–55	5	41	22	32	.290	.043	.001	12.20
68–73	—	6	42	52	.014	.004	.001	5.30
84–89	—	2	22	76	.004	.001	.0005	2.68
M-58-50								
0–2	—	88	8	4	.304	.230	.145	1.52
3–9	—	28	32	40	.074	007	.002	6.40
10–15	3	88	7	2	.270	.170	.115	1.53
17–21	2	16	24	58	.017	.002	.001	5.10
30–34	—	4	24	72	.004	.001	.0003	3.42
M-58-51								
0–3	—	90	5	5	.320	.230	.145	1.48
5–9	—	11	48	41	.028	.006	.002	4.00
20–25	—	9	41	50	.013	.004	.001	3.40
30–35	—	8	38	54	.011	.003	.001	3.40
40–47	14	69	7	10	.700	.360	.150	2.16
50–55	—	3	26	71	.005	.001	.0005	3.30
65–69	—	—	29	71	.005	.001	.0003	4.15
M-58-52								
0–3	—	83	14	3	.330	.240	.110	1.73
10–20	—	7	39	54	.023	.003	.001	2.70
32–35	1	74	14	11	.350	.240	.062	2.42

TABLE 4—*Continued*

Sample number	Gravel	Sand	Silt	Clay	Q_3 (mm)	Md (mm)	Q_2 (mm)	So
M-58-53								
0–3	—	88	5	7	.340	.250	.150	1.50
10–20	—	7	37	56	.009	.003	.001	2.60
32–25	—	61	16	23	.290	.125	.005	7.30
45–50	—	21	22	57	.025	.005	.0003	8.60
56–59	7	61	18	14	.300	.150	.030	3.16
60–65	6	53	21	20	.310	.115	.008	6.00
M-58-54								
0–5	—	80	11	9	.310	.200	.086	1.90
10–15	—	10	40	50	.017	.004	.001	4.10
40–45	—	4	34	62	.009	.002	.001	3.64
65–70	—	3	28	69	.005	.002	.001	2.36
95–102	—	3	34	63	.008	.002	.001	3.84
M-58-55								
0–10	—	2	60	38	.014	.007	.002	2.78
30–40	—	4	62	34	.019	.008	.002	2.82
50–60	—	2	63	35	.014	.006	.002	2.36
70–84	—	3	59	38	.015	.006	.002	2.73
M-58-56								
0–5	1	82	17	—	.280	.190	.088	1.78
5–15	—	11	42	47	.015	.005	.002	2.96
20–30	—	5	35	60	.008	.003	.001	2.54
40–50	—	2	29	69	.005	.002	.001	2.23
70–80	—	11	37	52	.015	.004	.001	4.60
90–95	—	24	49	27	.061	.016	.003	4.36
96–100	—	11	50	39	.027	.008	.001	4.50
M-58-57								
0–10	—	82	14	4	.134	.105	.072	1.36
10–16	—	69	25	6	.145	.088	.045	1.78
20–25	—	33	41	26	.074	.024	.003	4.65
M-58-58								
0–3	—	34	50	16	.145	.037	.008	4.24
5–8	—	79	17	4	.270	.158	.070	1.96
15–25	—	4	57	39	.012	.007	.002	2.64
40–50	—	2	49	49	.010	.004	.001	2.70
60–70	—	2	43	55	.008	.003	.002	2.27
70–84	—	1	48	51	.008	.004	.001	2.40
M-58-61								
0–5	2	73	18	7	.270	.140	.052	2.28
9–15	—	11	63	26	.031	.013	.004	2.92
30–40	—	5	61	34	.019	.008	.002	2.80
50–60	—	3	53	44	.012	.005	.001	2.93
75–86	—	2	44	54	.009	.003	.001	2.68
M-58-62	4	96	—	—	.560	.480	.370	1.22

microscope. The purpose of this examination was to observe the roundness and sphericity of the material, also to note the approximate percentage of pyrite, shells, and shell fragments present. Binocular examinations of 141 samples were made (table 6).

Heavy and Light Mineral Analysis

The material of sand size was shaken through Tyler sieves by means of a Ro-tap machine and the material within the range of .250–.125 mm was removed. This material was then passed through a micro sample splitter until approximately 3 grams of sample remained. The heavy minerals were separated by treating the sample with tetrabromethane (sp. gr. 2.95) using the method described by Krumbein and Pettijohn

TABLE 5.—*Percentage mechanical composition and statistical constants of late Wisconsin till samples*

Sample number	Gravel	Sand	Silt	Clay	Q_3 (mm)	Md (mm)	Q_2 (mm)	So
M-58-T1	9	67	6	18	.44	.260	.087	2.2
M-58-T3	8	61	9	22	.37	.210	.012	5.6
T-1	5	72	8	15	.37	.220	.074	2.2
T-2	15	70	4	17	.42	.270	.120	1.9
T-3	6	77	3	14	.45	.300	.150	1.7
T-4	3	67	6	24	.36	.220	.006	8.5
T-5	18	38	23	21	.40	.220	.01	6.3
T-7	8	65	8	19	.37	.300	.02	4.7
T-9	1	50	19	30	.32	.078	.002	11.
T-10	16	74	3	7	.62	.390	.230	1.6
T-15	2	72	10	16	.39	.250	.060	2 5
T-16	8	78	4	10	.44	.30	.170	1.6
T-18	4	63	15	18	.37	.220	.012	5.6
T-20	3	74	7	16	.40	.250	.088	2.1
T-22	5	61	19	15	.29	.170	.005	4.5
T-23	6	71	6	17	.38	.250	.10	1.9
T-24	2	70	8	20	.35	.220	.034	3.2
T-28	3	77	4	16	.39	.260	.13	1.7
T-30	3	62	15	20	.33	.162	.01	5.7
T-33	5	62	11	22	.35	.190	.008	6.6
T-36	11	71	4	14	.45	.280	.155	1.7
T-40	9	75	5	11	.31	.210	.150	1.4

(1938, p. 343). Identification of the heavy minerals was made with a petrographic microscope. The count of heavy minerals was made with a petrographic microscope equipped with a mechanical stage. Approximately 300 grains were counted in each sample.

The light minerals were treated according to the staining method described by Hays and Klugman (1959). The minerals were then identified and counted under a binocular microscope.

The light and heavy mineral composition of 6 samples was determined in this study (table 8).

Carbonate Content

Carbonate analyses were obtained for 141 lake samples and 20 glacial till samples (tables 6 and 7). The samples were shaken through a 1 mm sieve to remove the very coarse fraction. The samples were analyzed using the following method: representative portions were split from the bulk, sieved through a 1 mm screen, and ground approximately 2 minutes in a mortar. The carbonate content was then determined by means of a Chittick gasometric apparatus using 5N sulfuric acid and 1.7 grams of material in each case. The sample was digested in sulfuric acid and the volume of gas evolved was measured. This volume was corrected to standard temperature and pressure, and converted to percent calcite and dolomite per gram of sample (Lepper, 1950, pp. 118–119).

Clay Mineral Analysis

The clay mineral content of the less than .002 mm fraction was determined for 92 lake samples and 21 glacial till samples (table 9 and 10), using a G. E. XRD-3 X-ray recording diffractometer. Oriented aggregate slides were prepared by pouring a mixture of clay and distilled water on glass slides and allowing them to air dry. X-ray patterns were obtained both before and after placing the samples in an ethylene glycol desiccator for twenty-four hours. Samples were also X-rayed after heating to 450° C for a one hour period. In addition the less than .030 mm fraction was also studied. Differential thermal analyses of the less than .002 mm fraction were made for 6 samples. This combination of procedures made it possible to identify the types of clay minerals present in the samples (Grim, 1953; Weaver, 1956; Johns, Grim, and Bradley, 1954). Semi-quantitative estimates of the percentages of the principal clay minerals were obtained by

TABLE 6.—*Depth of water, percentage weight carbonate (<1 mm) and abundance of sand fraction shell fragments, shells, and pyrite (sample intervals given in centimeters for M-58-37 to M-58-61)*

Sample Number	Depth of water (ft.)	Percent carbonate	Shell fragments	Shells	Pyrite
M-58-1	35	6	P		
M-58-2	250	7			
M-58-4	250	22			
M-58-6	45	1			
M-58-7	60	4			
M-58-8	43	0			
M-58-9	60	0	A	A	
M-58-10	73	0			
M-58-11	22	0			
M-58-12	42	0	P	P	
M-58-13	52	0	P	P	
M-58-14	52	5			
M-58-15	75	0			
M-58-16	113	0			
M-58-17	100	0			
M-58-18	100	4		C	
M-58-19	145				
M-58-20	175	3			
M-58-21	150				
M-58-22	140	1	P		
M-58-23	150	2	P		
M-58-24	150	0			
M-58-25	150	0			
M-58-26	130	2	P	P	
M-58-27	150	8			
M-58-28	175	5			
M-58-29	100	1			
M-58-30	150	0			
M-58-31	180	14			
M-58-32	180	13			
M-58-33	150	17			
M-58-34	120	0			
M-58-35	120	0			
M-58-36	170	0			
M-58-37	275				
0–6		37	C	C	
10–15		22			
20–25		22			
30–34		26			
40–45		30			
50–54		12			
M-58-38	400				
0–8	12	12	P		
20–30		16			
50–60		16			
75–85		19			
M-58-39	400				
0–3		4			
10–20		16			
40–50		19			
60–70		18			
80–88		18			
M-58-40	250				
0–3		0			
10–20		44	C		

P = present (less than 1%); C = common (less than 10%); A = abundant (greater than 10%).

TABLE 6—*Continued*

Sample Number	Depth of water (ft.)	Percent carbonate	Shell fragments	Shells	Pyrite
35–40		55	C		
42–47		33	A	A	
48–53		31	P		
53–56		27	P		
M-58-42	225				
0–2		0			
5–10		45			
12–15		26	P		
M-58-43	400				
0–10		11			
20–30		11			
50–60		17			
75–87		16			
M-58-44	193				
0–10		20			
30–40		18			
50–60		20			
75–82		26			
M-58-45	180				
0–10		16			
30–40		16			
50–60		21			
75–83		29			
M-58-46	150				
0–3		41			
20–30		44			
50–60		51	P		
70–80		53	C		
95–98		32	A	C	
100–105		28	P		
106–109		20			
M-58-47	120	1			
M-58-49	150				
0–3		1			
20–30		46	A	A	
50–60		49	A		A
61–65		14	C	C	P
68–73		28	P		
84–89		44	P		P
M-58-50	165				
0–3		0			
3–9		27	C		
10–15		1			
17–21		29			
30–34		42	P		
M-58-51	170				
0–3		0			
5–9		27			
20–25		44	P	C	
30–35		40	C	C	
40–47		11	C		
50–55		34			
65–69		55			

TABLE 6—*Continued*

Sample Number	Depth of water (ft.)	Percent carbonate	Shell fragments	Shells	Pyrite
M-58-52	140				
0–3		2			
10–20		41	P		
32–35		9	A	A	
M-58-53	150				
0–3		1			
10–20		42	A		
32–35		8	A	A	
45–50		5	P		
56–59		11			A
60–65		15			
M-58-54	365				
0–3		8			
10–15		45	P		
40–45		46	C	C	A
65–70		51	C		A
95–102		34	C		A
M-58-55	340				
0–10		11			
30–40		12			
50–60		15			
70–84		13			
M-58-56	210				
0–5		0			
5–15		23			
20–30		40	P		A
40–50		51	A		A
70–80		47	A	A	A
90–95			A	A	
96–100		29	A	A	
M-58-57	175				
0–10		9			
10–16		18			
20–25		23			
M-58-58	115				
0–3		19			
5–8		5			
15–25		36			
40–50		41	C		
60–70		42	P		
70–84		41			
M-58-61	210				
0–5		2			
9–15		20			
30–40		29			
50–60		37			
75–86		41			
M-58-62	82	0			

a comparison of the areas contained beneath the basal (001) reflection peaks of the mineral, using the method of Johns, Grim, and Bradley (1954, pp. 242–251). These areas were multiplied by factors of 1 for chlorite and mixed layer minerals and by a factor of 2 for illite.

SEDIMENTS OF NORTHEASTERN LAKE MICHIGAN

Bottom Sediment Distribution

A map showing the general distribution of sediment types in relation to the bottom topography of northeastern Lake Michigan has been prepared (fig. 5). This map is based on size analyses of 50 samples taken from approximately 0–20 cm depth in the bottom.

The sediments are grouped in general according to the textural classification suggested by Shepard (1954). Five classes of sediment were designated: (1) Gravel-Sand-Silt-Clay, containing greater than 10% gravel and greater than 40% sand; (2) Sand, containing greater than 75% sand; (3) Clayey and Silty Sand, containing at least 45% but less than 75% sand; (4) Sandy "Clay," containing both sandy clay and sandy silt which have a sand content of 10–45%; (5) "Clay," containing both silty clay and clayey silt with less than 10% sand.

TABLE 7.—*Percentage weight carbonate of late Wisconsin till samples*

Sample	Percent carbonate
M-58-T1	8
M-58-T3	20
T-1	7
T-2	6
T-3	6
T-4	15
T-5	19
T-7	16
T-9	1
T-10	4
T-15	10
T-16	1
T-18	19
T-20	10
T-22	14
T-23	13
T-24	12
T-28	7
T-30	13
T-33	18
T-36	8
T-40	5

This map (fig. 5) clearly shows that the texture of the sediments does not decrease in size with increasing distance from shore. In the ridge and valley area (south west corner of fig. 5) the sediment texture ranges from sand and clayey sand covering the

TABLE 8.—*Percentage of heavy minerals in grade size .250–.125 mm. of northeastern Lake Michigan samples*

Sample Number	Pyrite	Magnetite	Weathered Minerals	Pyroxene	Amphibole	Epidote	Zircon	Garnet	Tourmaline	Total % Sillimanite Sphene Rutile
M-58-56 5–10 cm	1	12	71	2	11	1	1	1	0	0
M-58-56 20–25 cm	53	6	33	1	4	1	1	1	0	0
M-58-56 45–50 cm	83	9	1	0	4	0	1	1	0	1
M-58-56 70–76 cm	81	9	0	1	3	1	1	1	0	3
M-58-56 90–95 cm	1	9	33	5	40	2	3	5	0	2
M-58-2	2	15	33	2	28	5	1	13	0	1

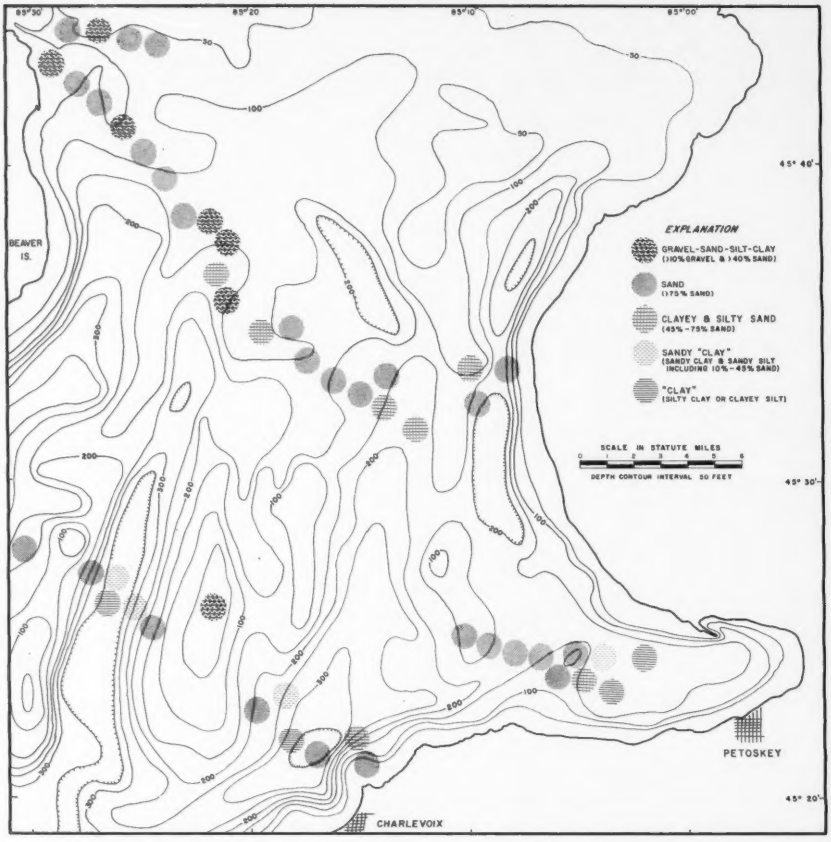

FIG. 5.—Areal distribution of sediments based on gravel, sand, silt, and clay content.

bottom close to shore, to "clay" and sandy "clay" four miles from shore, and gravel eight miles from shore. The sediment texture in the shelf areas (northwest and central portion of fig. 5) are less variable than those mentioned above, consisting in general of areas of sand and gravel. Little Traverse Bay contains areas of "clay" in the central part of the bay and a mixed zone of clayey and silty sand, and sandy clay at the western edge of the bay.

However, the sediment distribution is closely controlled by the topography and the depth of water. Topography is the primary factor in most of the area and especially in the ridge and valley portion of the map. The ridges and their adjacent flanks are covered with gravel and sand while the troughs are covered almost entirely with "clay" or sandy "clay." Coarse sediment is present on the ridges and flanks regardless of their distance from shore or depth of water. The texture of sediments in the shelf areas is controlled by both depth of water and topography.

Relation of Particle Size to Depositional Environments

The direct relationship between bottom topography and textural variation of sedi-

ments has been observed by Trask (1932), Revelle and Shepard (1939), and Emery (1960) in studies of sediments off the coast of California. It was shown in these studies that the texture and other sediment characteristics are very closely related to the bottom configuration, and there was a lack of correlation between sediment texture and distance from shore. Results of the present study indicate that the sediment distribution in northeastern Lake Michigan is also closely controlled by the bottom topography. In order to show this relationship cross sections were constructed along sample traverse lines. These sections illustrate the variations of percentages of gravel, sand, silt, and clay and the statistical constants of size distribution in relation to the bottom configuration. The locations of the cross sections (fig. 6, 7, and 8) are indicated on figure 4. The data on which these sections are based are listed in table 4. These cross sections are discussed in terms of the different lake environments.

Ridge and Valley sediments.—The southern portion of the area of study lies within the ridge and valley province of Lake Michigan (figs. 1 and 5), and the line of samples taken along section A–A′ (fig. 4) is almost perpendicular to the north-south trend of the ridge and valley topography. This complex topography, as shown in figure 6, consists of three deep troughs 350 to 400 feet deep. These troughs are separated by ridges covered by 25 to 100 feet of water. The northeast-southwest trend of the steep slope off Charlevoix is formed by the Petoskey Limestone. Samples taken along section A–A′ were spaced at one mile intervals. The types of sediment found in this area are directly related to the bottom topography (fig. 6). The systematic variation of median diameters of samples taken at 0–10 cm and at 10–20 cm depths on the lake bottom is shown in figure 6. Sediments covering the ridges and their adjacent flanks have the greatest median diameters (5 mm) and those found in the troughs have the smallest median diameters (.0058 mm). Only one sample (6) was obtained from the crest of

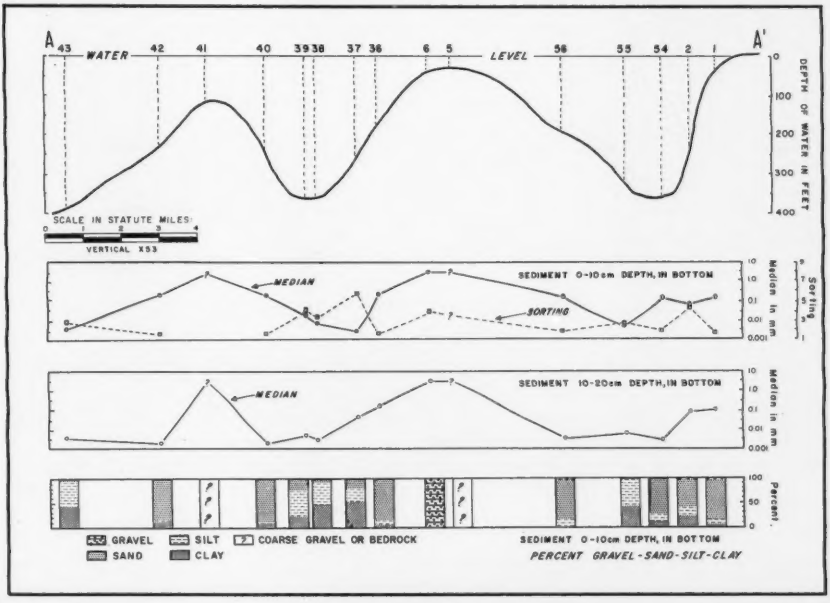

FIG. 6.—Ridge and valley sediments.

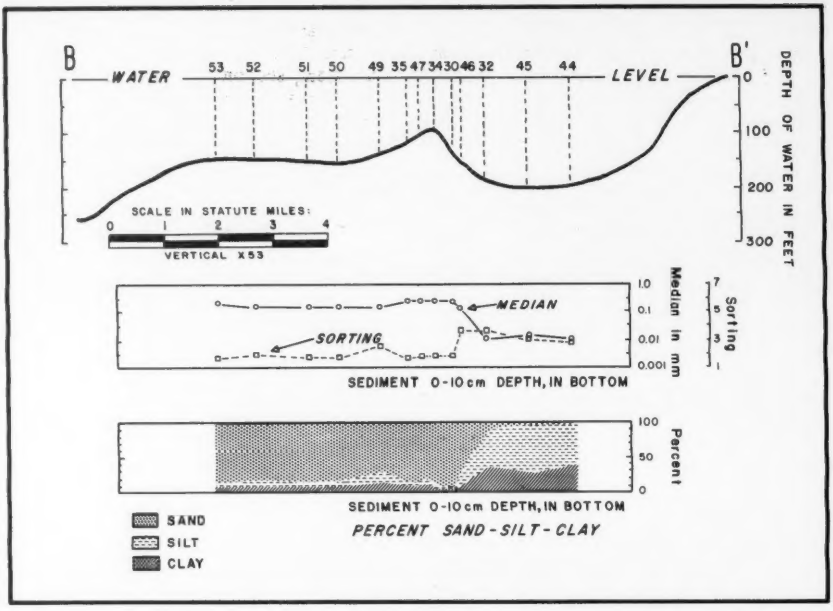

FIG. 7.—Little Traverse Bay and adjacent shelf sediments.

the ridges. Attempts to sample the ridges at locations 5 and 41 with the heavy coring tube and dredge failed. Therefore, it is probable that these topographic highs are either covered by a very coarse lag gravel deposit of the coarser constituents of till or are bedrock. The sediments obtained from upper flanks of troughs (36, 40, 42, and 56) consist primarily of sand having median diameters between .26 and .19 mm. It is interesting to note that samples taken at opposite sides of the flank in the same depths of water show a close correspondence in their median grain size. The majority of sediments in the deep water troughs are "clay" or sandy "clay" of a fine texture varying from .032 to .0058 mm. Sample 54, from the trough adjacent to the south shore of the area, contains a large percentage of sand and does not correspond to the usual trough sedimentation. This is probably due to its nearness to shore and the steepness of the adjacent flank.

The variation in median diameter of the sediment obtained at a 0–10 cm depth does not show the close correspondence with changes in topography as does the sediment obtained at 10–20 cm depth. This condition possibly indicates a lack of adjustment of the topographic profile with the more recent changes in lake level.

Most of the sediments in this environment are well sorted with values varying from 1.54 to 3. However, a few samples (2, 6, 37, and 39) have sorting values between 4 and 5.5. Samples 2 and 37, which were taken from the lower portions of the flanks, probably show poor sorting because of slumpage of coarse sediment from the ridges or their adjacent flanks.

It has been shown that wave action is most vigorous on the ridges during periods of high winds and this action would have been greatly accelerated during the low-water stage. The sediment distribution in this area reflects this action. Fine sediments have been winnowed away from the ridges and their side slopes, and deposited in the

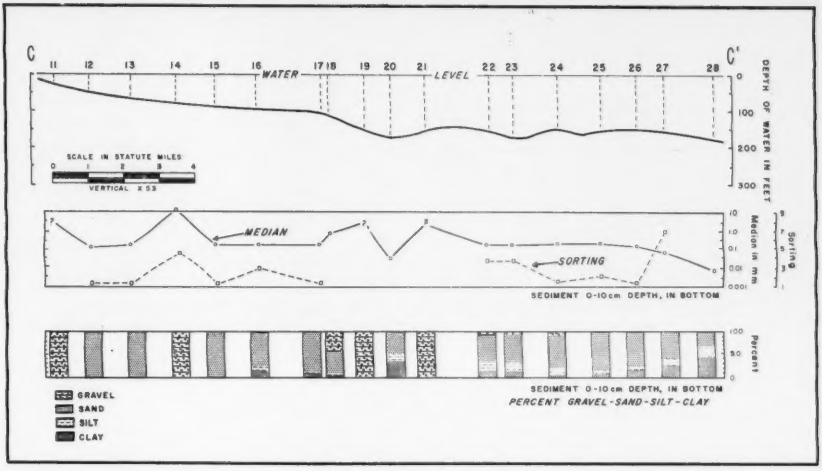

FIG. 8.—Shelf sediments.

deeper water of the troughs. Therefore, the pattern of valley and ridge sedimentation is closely controlled by the topography.

Bay and associated shelf sediments.—The line of samples taken, along section B–B′ (fig. 4), in Little Traverse Bay and the shelf area outside the bay is shown in figure 7. This small bay which has a maximum depth of less than 200 feet, is partially protected from the affects of both wave and current action by an elongated ridge at the western edge of the bay. The sediments of this environment show a regular variation with changes in depth of water and variation in the topography. The deposits covering the central portion of the bay have a maximum median diameter of .018 mm while those covering the elongate ridge and the broad shelf area west of the ridge have a maximum diameter of .260 mm. The sediments within each of these environments show little variation. Clayey and silty sediments were found in deep water of the central portion of the bay and sand size sediment predominates on the ridge and in the broad shelf area. The shelf sediments show good sorting while the bay sediments show poorer sorting. The regular occurrence of sandy deposits covering the shelf and ridge areas indicates that subaqueous erosion is very active in this area of the lake. The distribution of sediments in this environment appears to be related to both depth of water and topography.

Shelf sediments.—The northwest and central portion of the area lying north of the ridge and valley province can be characterized as a shelf because of the lack of variation in bottom topography. The line of samples taken along section C–C′ (fig. 4) in the shelf area east of Beaver Island is shown on figure 8. The lake topography in this shelf region shows little variation in relief, with a maximum depth of only 150 feet and an average of about 100 feet. Samples collected had medians between 13 mm and .009 mm which show a systematic variation in texture with the coarse gravel and sand sediments present in shallower water and fine sand or sand mixed with silt and clay covering the areas of deeper water. For example samples 25, 26, 27, and 28 show a gradual increase in silt and clay content with increasing depth of water. The sediments covering the shallow troughs in the shelf are somewhat finer than the other shelf sediments. Samples 2, 3, 24, and 25, which consist of sand and gravel mixed with a pink colored "clay," may represent either glacial till or a lag concentrate of the till.

The other sediments in this environment consist in general of well rounded gravel and cobble deposits, and sand. The coarse gravel deposits at locations 14, 19, and 21 may have been formed during the low water stage of the lake by subaerial erosion of the underlying till. Most of the sediments along this profile are well sorted. However, the samples taken on the flanks of the shallow troughs show poorer sorting values. This is probably the result of slumpage. The widespread occurrence of gravel and cobble, and sand deposits with very little clay or silt size material probably indicates that subaqueous erosion is very active in this shelf environment. The sediment distribution is closely related to the depth of water.

Sedimentary Sequence in Core Samples

The location of core samples taken in the ridge and valley area (A–A′) and in the Little Traverse Bay and adjacent shelf area (B–B′) is shown in figure 4. Five core samples were obtained in the deep troughs of the ridge and valley area from depths greater than 340 feet. Ten cores were taken in the shallower water on the flanks of the deep troughs and in the shallow water shelf area west of Little Traverse Bay in depths less than 275 feet. In addition, two cores were taken in the central portion of Little Traverse Bay in depths less than 200 feet, and one core was obtained at the western margin of the bay in 150 feet of water. The logs of these cores indicating the variations in the sediment texture (based on mechanical analyses), color, and organic content in relation to the lake bottom topography are shown in figures 9 and 10. The statistical constants and descriptions on which these logs are based are shown in table 4.

The sequence of sediments in these cores shows a definite relationship to the depth of water and the bottom topography. The logs of cores taken in deep water troughs display little vertical variation. On the other hand logs of cores taken on the flanks of the troughs and the shelf area show great variety in their lithology. In view of these variations, the cores taken in shallow-water in depths less than 275 feet, and the cores taken in deep-water troughs greater than 275 feet will be discussed separately. In addition, the three cores taken in Little Traverse Bay will be discussed as a separate group.

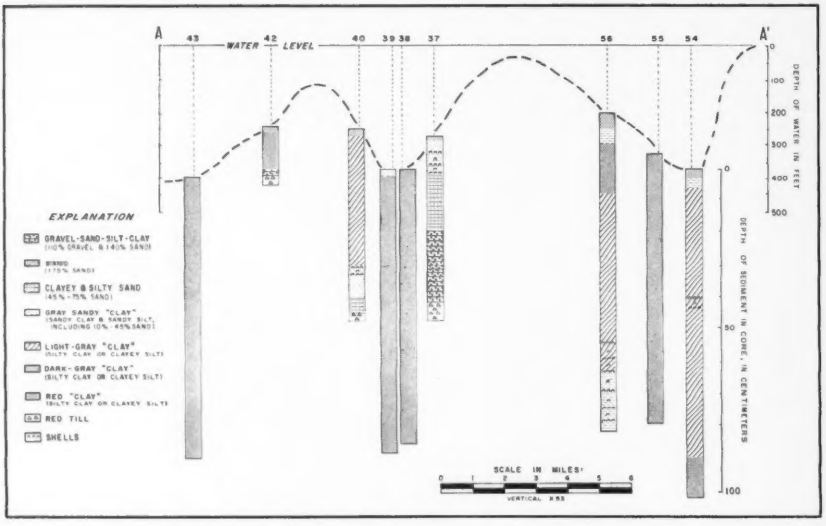

FIG. 9.—Lithology of core samples in ridge and valley area.

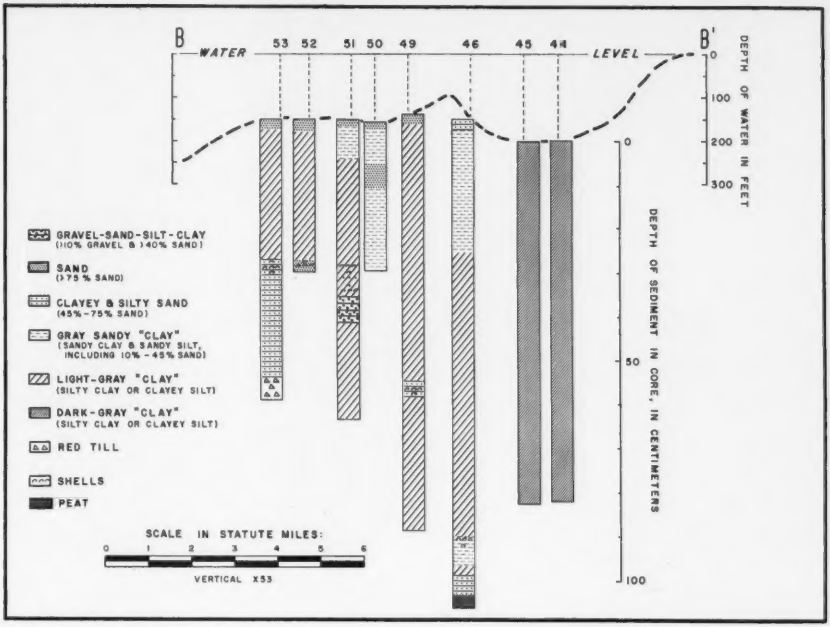

FIG. 10.—Lithology of core samples in Little Traverse Bay and on adjacent shelf.

Shallow-water core sequence.—The core samples (37, 40, 42, 49, 51, 52, 53, and 56) which were taken in "shallow-water" in depths of water between 150 and 275 feet show a great variation in their sequence of sediments (figs. 9 and 10). These variations are the result of differences in topography of the lake bottom, depth of water, depth of penetration of the core, and changes in lake level.

An idealized sequence of sediment zones found in shallow-water cores is shown below. This sequence is based on a correlation of these sediments with shallow water cores (depths less than 350 feet) taken in the central and southern parts of Lake Michigan (Hough, 1955).

Zone I—yellow sand, grading to
Zone II—light gray "clay"
Zone III—light gray sandy "clay" and/or light gray clayey and silty sand and/or gravel-sand-silt-clay
Zone IV—light gray "clay"
Zone V—red glacial till

Zone V—the red glacial till is present in cores 53, 42, 40, and 37 at a depth of approximately 60 cm in the cores. Its absence in the other shallow water cores is probably due to the fact that the core sampler did not penetrate far enough. The median diameters of samples from this zone show a very restricted range with a maximum diameter of .2 mm and a minimum of .115 mm. This unit contains 2–14% gravel, 42–59% sand, and 19–21% silt, and 16–25% clay. The sediment in this zone is poorly sorted, with values varying from 6.0 to 10. This very compact, poorly sorted, coarse-textured sediment, which is found at the bottom of some shallow water cores, correlates with the Valders age glacial till found on the Southern Peninsula of Michigan and with red till found in the central and southern parts of Lake Michigan (Snodgrass, 1952, pp. 22–23; Hough, 1935, p. 69; Hough,

1955, pp. 957–968). A comparison of this till zone with glacial till found on the adjacent land area will be presented later.

Zone IV—Overlying this red glacial till is a very fine grained light gray "clay" zone. This unit is present only in cores 51 and 49. Samples collected from this zone have median diameters ranging from .0013 to .004 mm. They show a range in particle size of 0–6% sand, 22–42% silt, and 52–57% clay.

Zone III—The sediments in this zone show the greatest variation in texture and stratigraphic position. This zone which is present in all shallow-water cores has median diameters with a range from a maximum of .360 to a minimum of .043 mm. The variation in texture of this light-gray coarse sediment zone is as follows: 0–17% gravel, 40–69% sand, 7–24% silt, and 10–32% clay. Following the classification used in this study the sediments in this zone are classed as gravel-sand-silt-clay, clayey and silty sand, and sandy "clay." The stratigraphic position of this unit is very variable. In some cores (37, 40, 42, and 53) this zone overlies the red glacial till, while in others (49 and 51) it overlies the finer grained, light-gray "clay" zone. The thickness of this unit varies from 2 to 40 cm. It is interesting to note that shells and shell fragments were present in this zone in cores 49, 52, and 53 and, as shown in figure 9, they are correlative between these cores. A sedimentary unit having essentially the same texture, shell content, color, and stratigraphic position has been noted in cores taken from less than 350 feet in the central and southern portion of the lake (Snodgrass, 1952; Hough, 1955).

Zone II—Overlying this coarse textured unit in cores 40, 49, 51, and 53 is a light-gray "clay." The sediment in this zone has a much finer texture than the Zone III sediment. The median diameters of these samples range from a maximum of .004 to a minimum of .0026 mm. There is little variation in the texture of this sediment. The range of particle size of the zone is as follows: 2–9% sand, 35–44% silt, and 50–62% clay. It should be noted that the sediments of this zone have the same textural and color characteristics as the samples analyzed from Zone IV.

Zone I—The uppermost sediment of all shallow-water cores is a yellow sand. The thickness of this unit ranges from 2 to 10 cm. The median diameters of samples from this zone show a very restricted range with a maximum of .250 mm and a minimum of .140 mm. This unit contains 76–92% sand, 3–17% silt, and 0–10% clay. The sediment in this zone is well sorted, with values varying from 1.48 to 2.34. It is interesting to note that these sediments show the best sorting values.

The sedimentary sequence in core 42 (fig. 9) differs slightly from the other shallow water cores. In this core a red "clay" zone occurs above a coarse red clayey gravely sand which in turn overlies red glacial till. No red "clay" zone is present in the other cores in this area. However, in a study by Hough (1955) of cores from southern and central Lake Michigan, a zone of this type was present overlying red glacial till.

Deep-water core sequence.—The core samples 38, 39, 43, 54, and 55 were taken in "deep-water" at depths of water between 340 and 400 feet. The deep-water cores show little variation in their lithologic sequence of sediments (fig. 9). The homogeneous sequence observed in these cores is a reflection of topographic position (deep-water troughs) at which they were taken and the depth of penetration of the core. Almost all these cores contain dark gray "clay" throughout the length of the cores (82 to 102 cm depth in bottom). The analyses of this very fine-grained dark gray sediment showed that the median diameter ranges from a maximum of .008 to a minimum of .0019 mm. The range of particle size for this material is as follows: 1–4% sand, 32–63% silt, and 34–69% clay. This deep-water dark gray "clay" displays a very close textural similarity to the light gray "clay" of Zone II and IV of the shallow-water sequence. Representative samples of dark-gray "clay" and light-gray "clay" having the same median grain size were heated (after removal of hygroscopic moisture) to 550° C for six hours. The dark gray "clay" had an average weight loss of 9.5% and the light-gray "clay" had an average loss of 4.3%. The loss of weight is progably due mainly to a loss of organic material. Therefore, this analysis

shows the dark-gray sediment to have approximately twice as much organic matter as the light-gray "clay." This increase in organic matter may indicate preservation of this material under reducing conditions and a greater amount of organic material deposited in the deep troughs. The sedimentary sequence in core 54 (fig. 9) differs from the other deep-water cores. This core contains an upper zone of sand overlying a light-gray "clay" zone and a bottom zone of dark-gray "clay." This variation from the normal deep water sequence is probably caused by both the proximity of this core sample to the shoreline and the steep slope of the trough.

Little Traverse Bay core sequence.—The core samples 45 and 44 that were taken in the central part of the bay in depths of water of 175–193 feet contained the same sequence of sediments. On the other hand core 46, taken in shallower water (150 feet) at the western margin of the bay, contained an entirely different lithology (fig. 10). The core samples from the central part of the bay contain a homogeneous sequence of fine grained dark-gray clayey silt. The analyses of this dark-gray sediment showed that the median diameter ranges from a maximum of .018 to a minimum of .007 mm. These cores show a range in particle size of 3–5% sand, 59–74% silt, and 23–36% clay. It is obvious that these sediments show a striking similarity in both color and texture to those cores taken in the deep-water of the valley and ridge area. The sedimentary sequence in core 46 shows a variation in lithology (fig. 10). The sediment in this core has essentially the same lithologic succession as the shallow-water cores. This core contains black peat at the bottom of the core overlain by clayey and silty sand grading to a light-gray "clay" which in turn grades to a light-gray clayey and silty sand at the top of the core. Analyses of the sedimentary units listed above show that they correlate in both texture and color with the sediments found in zones III, II, and I of the shallow-water sequence.

A detailed pollen analysis and interpretation of the environment of deposition for the peat sample from core 46 was mde by Mr. Russel Peppers. The peat contained 16 percent *Sphagnum* and *Typha* pollen which are derived from plants requiring standing water. The presence of *Sphagnum* spores shows that this environment was conducive to the formation of a bog peat, of an unknown size. The abundance of *Typha* (cattail) appears to indicate that the sample was taken from an area marginal to the bog. The high frequency (62%) of *Pinus* (pine) pollen probably indicates the climatic conditions for the formation of the peat were not far removed from those of the present. On the other hand, the absence of beech (*Fagus*) and maple (*Acer*) points to perhaps a slightly cooler climate than at present.

Interpretation of Post-Glacial Lake History from Core Samples

The relationship between shallow-water core zones, deep-water cores, and Little Traverse Bay cores is summarized in figure 11. It is obvious from this drawing that the texture of the sediment in Zone II and Zone IV of the shallow-water cores, in deep-water cores, and in cores from the central part of Little Traverse Bay are strikingly similar. The texture of Zones I, III, and V of the

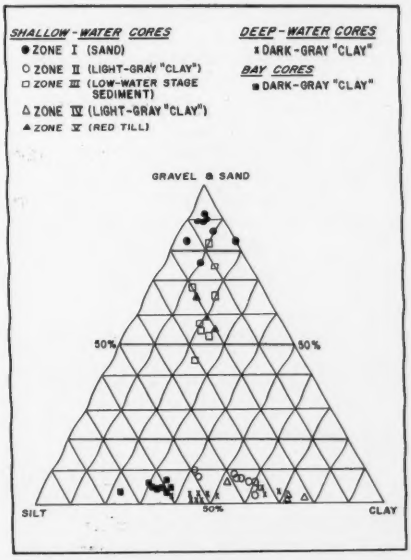

FIG. 11.—Percentages of gravel and sand, silt and clay in core samples.

shallow-water sequence is much coarser than that of the aforementioned sediment.

These textural differences are considered a valid means for interpretation of the post-glacial history of this northeastern portion of Lake Michigan. The sediments in Little Traverse Bay (except core 46) and those found in the deep-water troughs show an uninterrupted sequence of dark-gray "clay." On the other hand, the shallow-water sediments (depths less than 275 feet) contain a zone of very coarse sediment (Zone III) overlying, underlying, or included within a sequence of much finer grained light-gray "clay" (Zone II and IV). This light-gray "clay" has a texture similar to the deep-water sediments (fig. 11). The stratigraphic position of the very coarse sediment of Zone III is very variable: in some cores it overlies the red glacial till, in others it overlies the finer grained light-gray "clay" zone. The shallow-water sequence, therefore shows great variation in its sedimentary pattern, and these variations reflect the changing environment of the shallow water deposits.

The coarse texture of the material in Zone III which unconformably overlies glacial till in some cores and truncates light-gray "clay" in other cores is interpreted as representing sediment deposited in a shallow-water environment. This shallow- or low-water environment is probably correlative to the Lake Chippewa low-water stage of Lake Michigan (Hough, 1955). During this stage the lake level was 270 feet lower in relation to the bottom (in this area), therefore exposing the area in which these shallow-water cores (depths less than 275) were taken to wave action and subaerial erosion. The light-gray "clay" (Zone IV) overlying the Valders till was probably removed by erosion during this stage and replaced by coarser material. The light-gray "clay" overlying the low-water stage sediment (Zone III) represents the resumption of deep-water conditions which correspond to post-Chippewa to Nippissing time (fig 2). The sequence of coarse gray sediment (Zone III) overlying glacial till in some cores is probably evidence of an unconformity because of the following: sharp contact between the two zones, differences in color of sediments, and change in texture. The deposit of sand in the top 10 cm of the shallow water cores may be caused by the lowering of lake water 35 feet from the Nippissing stage (fig. 2).

A partial history of post-glacial sedimentation in Little Traverse Bay can be observed in cores 44, 46, and 45 (fig. 10). During the Chippewa low-water stage this bay was probably covered by only a few feet of water as indicated by the peat material found at a depth of 110 cm in core 46. Overlying this peat and underlying a light-gray "clay" (texture similar to Zone II) is found a sequence of gray sandy "clay" with shells. This gradational zone, as shown in figure 10, shows progressive deepening of water following the Chippewa low-water stage. The cores from the central part of the bay, on the other hand, show a marked similarity in lithologic sequence, color, with those taken in the deep-water troughs. It should be noted, however, that the dark-gray "clay" from Little Traverse Bay contains a higher percentage of silt.

Additional evidence of the changing sedimentary environment in this area is shown by the presence of shells of a shallow water fauna in the lake cores (figs. 9 and 10). The pelecypod and gastropod shells were identified by Professor Max Matteson as forms which live in extremely shallow open-lake water. Three types of aquatic gastropods were recovered from the cores. The most abundant gastropod was *Valvata tricarinata* which commonly lives in less than 6 feet of water. The genus *Aminicola* which lives in 1–15 feet of water and the genus *Helisoma* which lives in quiet water 1–9 feet deep were also present. Pelecypods of the family Sphaeriidae were very abundant. Most of these pelecypods are referred to the genus *Sphaerium* which lives in water less than 3 feet deep where the current action is strong. The shells identified in this study are essentially the same as those found by Hough (1955, p. 961) in a study of shallow-water cores from central Lake Michigan. In general the shell zones are present in, or are closely related to the coarse sediment of Zone III.

The sediments found in the shallow-water cores (37, 46, 49, 56) show evidence of progressive deepening of lake water. These cores show a transitional zone of gray sandy

"clay" (sometimes with shells or shell fragments) in the core sequence. In some cores this zone overlies the coarse textured sediment of the low-water stage.

Hough (1955) has shown a very similar sequence of sediments in shallow-water cores from central Lake Michigan. In this study, he observed a thin zone of sand (some with shells of shallow-water forms). This zone was included between a sequence of finer grained red or gray deep-water clay. The mechanical analyses of this sand zone and other lithologic units in these shallow-water cores showed a very close correspondence in both texture and color with the shallow-water cores analyzed in this study (Snodgrass, 1952; Hough, 1955).

Carbonate Content of Lake Sediments

Calcite and dolomite are present in the sand fraction of the lake sediments. Most of the carbonate occurs as grains of limestone and dolomite, and only a very small proportion is present as shells and shell fragments. A comparison of the carbonate content of samples containing shells to material of the same texture and color without shells showed little increase in the carbonate percentage. Dolomite is present in the silt and clay fraction (X-ray analysis) of most samples and calcium carbonate is present in a few samples.

The relationship of median diameter and carbonate percentage in open lake samples (excluding till and low-water stage sediment) is plotted on figure 12. The carbonate content of the samples shows a direct relationship to the sediment size. In general the carbonate content increases with a decrease in the median diameter of the sediment. The samples from the ridges show the smallest percentage of carbonate, and the flank and some shelf samples show the highest carbonate percentage. The reduction of carbonate in these coarse-grained samples is probably due both to solution and to grain size reduction as a result of abrasion, followed by winnowing out of the finer material. Dredge samples (11, 19, 21) taken along traverse C–C′ (fig. 8) contain limestone cobbles with numerous solution pits. There is some indication from these data that the variation of carbonate content is in part related to the distance from the source area

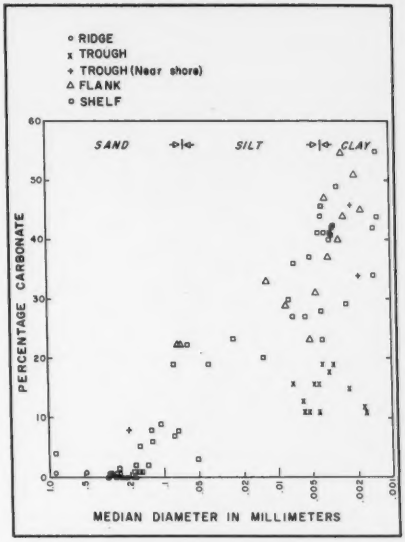

FIG. 12.—Relationship of median diameter and carbonate content in open lake sediments (excluding low-water stage sediment and glacial till).

and the rate of sedimentation. Sediment present in core 54 which is in a deep trough near shore contains a thick sequence of light-gray "clay" with an average carbonate content of 47%. On the other hand sediment in core 55 which is present in the same deep trough but at a greater distance from the source area (shore or flank of trough) contains only dark "clay" with an average carbonate content of 13%. The association of higher carbonate values with finer sediments has been noted by Caldwell (1940, pp. 58–64), Shepard and Moore (1954, pp. 1463–1593).

The relationship of median diameter and carbonate content in core samples from the open lake water and Little Traverse Bay is shown in figure 13. The carbonate content of sediment from the light-gray "clay" (Zone II and Zone IV) of the shallow-water cores showed the highest percentage of carbonate with values ranging from 26% to 55%. The material having the lowest percentage of carbonate is from the sandy (Zone I) material found in the upper 10 cm

of the cores. Intermediate between these two sediments were the dark-gray "clay" samples from the deep water troughs with values ranging from 11–39% and the dark-gray "clay" from Little Traverse Bay. The carbonate content of the low-water sediment (Zone III) showed the widest range in percent carbonate and a close similarity between this zone and the red till (Zone V) was noted.

The ranges of percent carbonate in the dark-gray "clay" samples from the deep troughs and from Little Traverse Bay are distinct as is shown by the dashed line enclosing these sediments in figure 13. Light-gray "clay" sediment from shallow-water which has approximately the same median diameter as those noted above has much higher values of carbonate. The decrease in carbonate content in the dark-gray "clay" found in the deep troughs and bay suggests that carbonate has been removed from these sediments either during transport or after deposition. Analyses of the dark-gray "clay" showed a consistently lower percentage of carbonate than did the light-gray "clay". The dark-gray "clay" has an average carbonate content of 15% for samples from deep-water troughs and an average of 21% for cores from Little Traverse Bay. On the other hand the light-gray "clay" which has approximately the same median diameter has an average carbonate content of 43%. X-ray analyses of the silt (less than .03 mm) and clay (less than .002 mm) fractions of these sediments revealed the following: the dark-gray "clay" contained dolomite but no calcite, and the light-gray "clay" contained both calcite and dolomite in all samples (tables 6 and 9). Carbonate and X-ray analyses of the Valders glacial till (source material) on the adcent shore showed the presence of both calcite and dolomite in the silt and clay fraction of the till and the carbonate percentages ranged from 5–20% (tables 7 and 10).

The decrease of carbonate content and absence of calcite in the dark-gray "clay" sediments of the deep-water troughs and bay appears to be directly related to environmental factors. Allgeier, Hafford, and Juday (1941) who studied three oligotrophic lakes in Wisconsin showed that the hydrogen ion concentration decreased markedly with

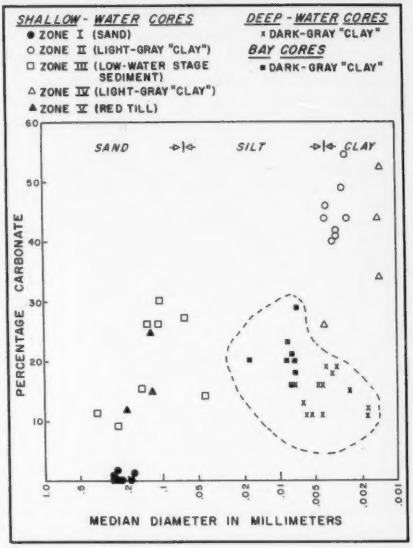

FIG. 13.—Relationship of median diameter and carbonate content in core samples from open lake and Little Traverse Bay.

increasing depth of water and in Crystal Lake the pH ranged from 7.1 at the surface to 5.9 at the bottom. No analyses have been made of the change of hydrogen ion concentration with depth in Lake Michigan, however, conditions are probably similar to that noted above. The decrease in pH in the dark-gray "clay" sediments could be brought about by decay of organic matter which would produce carbon dioxide and by concentration of this carbon dioxide in the cold stagnant water of the deep trough and bay. This lowered pH could remove calcite from the sediment and leave behind the less-soluble dolomite. The removal of calcite would therefore lower the total carbonate percentage of the sediments from the deep troughs and Little Traverse Bay.

It should be noted, however, that the decrease in carbonate percentage may also be accounted for in part by removal of this material during transport. In general it was noted that the sediment from "shallow" water cores showed higher carbonate percentages than did the deep-water cores (except 54). This suggests that the distance

from the source (topographic highs or shore) or distance of transport may also affect the carbonate percentage.

Relation of Lake Deposits to Source Materials

The source of sedimentary materials found in northeastern Lake Michigan is chiefly glacial drift deposits. These deposits include the Wisconsin age till found on adjacent land areas and on the lake bottom. The sediments that cover the deep troughs and the flanks of these troughs are probably derived for the most part from wave erosion of glacial till that underlies the ridges.

The statistical constants of size distribution of twenty-two glacial till samples is given in table 5. The till samples may be characterized as having a very high sand content with only a very small percentage of gravel, silt, and clay. The average median diameter of the till is .218 mm. The glacial till is poorly sorted with an average sorting of 3.8.

The carbonate content of both the Valders and Port Huron till is highly variable with values ranging from 1–20% (table 7). Binocular examination of the pebble lithology and X-ray analyses of the silt and clay fraction indicates that the carbonate fraction of the till contains both calcite and dolomite. The carbonate content of the Valders till is generally much higher than that of lake samples having the same median grain size.

The red sediment found at the bottom of cores 37, 42, and 53 show very definite correlation to the Valders age till found on the adjacent shore. The correlation between the two sediments is as follows:

(1) The two sediments show the same textural characteristics.
(2) Both sediments are poorly sorted, the till material in lake cores have sorting values ranging from 6–10.
(3) The carbonate content of the two sediments is essentially the same and both contain dolomite and calcite.
(4) Both materials exhibit the same red to pinkish-red color.
(5) The heavy mineral content of the two sediments are almost identical. Pyrite is absent from both sediments.

Diagenetic Alteration of Lake Sediments

A comparison of the heavy mineral content of the lake sediments and glacial till samples from the adjacent land areas revealed a striking variation in the type of iron minerals present. The mineral suite of thirty-six glacial till samples (Valders and Port Huron age till) was described by the author in a study of late Wisconsin tills in the northern part of the southern peninsula of Michigan (Moore, 1958). The predominant heavy minerals in these tills were magnetite, ilmenite, weathered minerals, hornblende, and garnet with moderate amounts of epidote and augite. The heavy mineral content of the lake sediments (seven samples) is similar and the predominant heavy minerals are iron minerals, weathered minerals, hornblende, and garnet. However, there is one notable exception: three samples from core 54 showed a pyrite content which was greater than 90% of the total metal opaques and comprised 54–83% of the total heavy mineral content (table 8). In addition, the examination of the sand fraction of all lake samples (141) revealed the presence of zones containing abundant pyrite (greater than 20% of the sand fraction). These zones were found in cores 49, 53, 54, and 56 (table 6). The iron sulfide is in the form of yellow muriform masses. The octahedral form of some of these grains indicate that the material is pyrite rather than marcasite (Emery, 1960, p. 267). Pyrite was not found in the glacial till zones in the cores or in any of the glacial till deposits from the adjacent shore.

These pyrite zones show the following relation to the lake sediments:

(1) Present within a restricted interval of 45–80 cm (except core 54).
(2) Always found in association with shells or shell fragments.
(3) Sediment in the cores with pyrite zones is light-gray sandy "clay" or "clay."

Pyrite was also found within the shells of pelecypods in cores 54 and 56, and many grains of pyrite showed inclusions of shell fragments. This very close association between pyrite zones and calcareous shell material strongly suggests a diagenetic ori-

TABLE 9.—*Clay mineral (<.002 mm), and calcite-dolomite (<.03 and <.002 mm) composition of northeastern Lake Michigan samples (sample intervals given in centimeters for M-58-37 to M-58-61)*

Sample number	Illite	Mixed layer	Chlorite	Calcite <.03	Dolomite <.03	Calcite <.002	Dolomite <.002
M-58-2	50	30	20			—	—
M-58-4	40	30	30			P	—
M-58-16	30	50	20			—	—
M-58-17	50	30	20			—	—
M-58-20	40	30	30			—	—
M-58-22	40	40	20			P	—
M-58-23	30	50	20			—	—
M-58-26	50	40	10			—	—
M-58-28	40	40	20			—	P
M-58-31	50	40	10			—	—
M-58-32	40	30	30			—	—
M-58-37							
0–6	30	40	30	P	P	P	P
10–15	50	40	10	P	P	P	P
20–25	60	20	20	P	P	P	P
40–45	70	20	10	P	P	P	P
50–55	40	40	20	P	P	—	—
M-58-38							
0–8	50	30	20	—	P		—
20–30	60	30	10	—	P	—	—
50–60	60	30	10	—	P	—	P
75–85	60	20	20	—	P	—	—
M-58-39							
0–3	60	30	10	—	P	—	—
10–20	50	40	10	—	P	—	P
40–50	40	30	30	—	P	—	P
60–70	40	40	20	—	P	—	P
M-58-40							
10–20	40	40	20	P	P	P	P
35–30	60	20	20	P	P	P	P
42–47	60	30	10	P	P	P	P
M-58-43							
0–10	50	30	20	—	P	—	P
50–60	50	30	20	—	P	—	P
75–87	50	30	20	—	P	—	P
M-58-44							
0–10	50	20	30	—	P	—	P
30–40	70	20	10	—	P	—	P
50–60	70	20	10	—	P	—	P
75–82	60	20	20	—	P	—	P
M-58-45							
0–10	40	40	20	—	P	—	P
30–40	60	30	10	—	P	—	P
50–60	70	20	10	—	P	—	P
75–83	50	30	20	—	P	—	P
M-58-46							
20–30	40	40	20	P	P	P	P
50–60	40	40	20	P	P	P	P
70–80	30	50	20	P	P	P	P
95–98	40	40	20	P	P	P	P
100–105	40	40	20	P	P	P	P
M-58-49							
0–3	30	50	20	P	P	—	—
20–30	50	30	20	P	P	P	P
55–61	40	40	20	P	P	—	P
62–65	30	50	20	P	P	—	P
84–89	80	10	10	P	P	P	P

P = present; — = absent

TABLE 9—*Continued*

Sample number	Illite	Mixed layer	Chlorite	Calcite <.03	Dolomite <.03	Calcite <.002	Dolomite <.002
M-58-50							
3–9	60	30	10	P	P	P	P
17–21	40	30	30	P	P	P	P
30–34	60	30	10	P	P	P	P
M-58-51							
5–9	40	20	40	P	P	P	P
20–25	30	40	30	P	P	P	P
35–40	30	40	30	P	P	P	P
50–55	40	30	30	P	P	—	P
M-58-53							
10–20	40	40	20	P	P	P	P
32–35	30	40	30	P	P	P	P
45–50	60	30	10	—	P	P	P
56–59	40	40	20	—	P	P	P
60–65	30	50	20	P	P	P	P
M-58-54							
10–15	40	50	10	P	P	P	P
40–45	50	40	10	P	P	P	P
65–70	40	40	20	P	P	P	P
95–102	40	30	30	P	P	P	P
M-58-55							
0–10	40	30	30	—	P	—	—
30–40	60	30	10	—	P	—	—
50–60	40	50	10	—	P	—	P
70–84	30	40	30	—	P	—	P
M-58-56							
5–15	50	40	10	P	P	P	P
20–30	50	40	10	P	P	P	P
40–50	40	50	10	P	P	P	P
70–80	40	40	20	P	P	P	P
96–100	60	30	10	P	P	P	P

TABLE 10.—*Clay mineral (<.002 mm), and calcite-dolomite (<.03 and <.002 mm) composition of late Wisconsin till samples*

Sample number	Illite	Mixed layer	Chlorite	Calcite <.03	Dolomite <.03	Calcite <.002	Dolomite <.002
T-1	30	20	50	P	P	—	P
T-3	50	20	30	P	P	P	P
T-4	40	20	40	P	P	P	P
T-5	60	30	10	P	P	P	P
T-7	40	40	20			P	P
T-9	50	30	20			—	—
T-10	60	20	20			—	—
T-16	60	10	30			—	—
T-18	60	20	20			P	P
T-22	50	30	20			P	P
T-23	50	30	20			P	P
T-29	30	20	50	P	P	—	P
T-32	60	30	10			P	P
T-35	40	40	20	P	P	P	P
T-30	60	20	20	P	P	P	P
M-58-T3	20	40	40			P	P

p = present; — = absent

gin for the pyrite. Local spots of negative Eh, at or near the depositional interface, could have been produced by the decay of soft parts of these shelled organisms. A marked decrease of Eh at these spots would cause pyritization. Krumbein and Garrels (1952) and Emery (1960) show that iron sulfide is formed under the same conditions in Recent marine sediments.

Clay Mineralogy

Clay mineral analyses of the clay size fraction (<.002 mm) were made for samples taken in deep troughs, flanks of troughs, shelf areas, and Little Traverse Bay. In addition glacial till samples from the adjacent shore were analyzed. The results of these analyses showed that the general order of abundance of the clay minerals in the samples was: illite, mixed-layer materials (undifferentiated), and chlorite. Illite was identified as a constituent of all samples by its 10Å peak. This first-order peak of illite was commonly asymmetrical to the low angles. The identification of the mixed-layer minerals was very difficult. In most samples the X-ray patterns of mixed-layer materials showed as broad shoulders held above background between 10.1Å and 14Å which was closely associated with first-order chlorite and illite peaks. In general the glycol treated material showed no expansion, although a few samples displayed a low-intensity shoulder between 14Å and 17Å positions. These mixed-layer clays are composed primarily of random interlaminated degraded illite, degraded chlorite, and chlorite. Possibly some mixed-layer chlorite, vermiculite, and montmorillonite is present. Chlorite was identified by its diagnostic 14Å reflection. The first order peak of chlorite was usually retained after heating to 450° C but the 7Å peak was either greatly diminished in size or lost. Poorly crystallized chlorite which was present in many samples was identified by its strong first-order reflection (14Å) and a weak second order reflection (7Å). Differential thermal analyses of four samples showed an exothermic reaction at 850° C. This corresponds to the formation of forsterite which is a high-temperature phase of chlorite.

The quantitative determinations of the different clay mineral groups in lake sediments are shown in table 9. The average clay mineral composition for all samples is as follows: illite 50%, mixed-layer minerals 30%, and chlorite 20%. A comparison of the average clay mineral composition of the lake sediments and glacial till from the adjacent shore shows little variation in the type of mineral present or in their relative abundance. The results of the quantitative analysis of Valders (11 samples) and Port Huron (5 samples) age till showed the following average clay mineral composition: illite 50%, mixed-layer minerals 30%, and chlorite 20% (table 10). In general these analyses show that the clay mineral content of the lake sediments reflects the composition of the source material.

A definite change in the clay mineral distribution was noted in a few samples. Some dredge and core samples taken in the shallow water shelf areas have a higher proportion of mixed-layer materials than illite (16, 22, 23, 33, 37, 46, 49, 51). In addition many of these same samples also show a much higher content of poorly crystallized chlorite. However, analyses of samples taken in the valley and ridge area and in the middle of Little Traverse Bay showed little variation in clay mineral content.

The observed differences in composition noted above are considered to be related to the texture of the sediment, weathering during transport, and weathering in place. In many cases the clay fraction of the sandy "clay," silty and clayey sand, and sand zones present in the shelf environment contain a high proportion of mixed-lattice minerals and poorly crystallized chlorite. These coarser grained zones probably would be more susceptible to subaqueous weathering action by waves and currents. This weathering could result in the degradation of chlorite and illite by the partial leaching of potassium and magnesium from the clay minerals. On the other hand sediments found in deeper water of the valley and ridge areas would probably not be as subject to this weathering action.

Several authors have shown that land-derived clays deposited in the fresh water lake environment will consist of illite and mixed-layer clays. According to Millot (Grim, 1953) the dominant mineral in lake sediments is illite when there is much calcium carbonate present. Studies by Cuth-

bert (1944) of 14 samples from Lake Erie showed that illite and mixed-layer material was present, with illite as the dominant clay mineral. He points out that the clay mineral composition of these sediments is probably a reflection of the source material. A recent study by Rogers (1959) who analyzed 8 samples of deep water sediment from northwestern Lake Huron showed the sediments were composed of chlorite, illite, and montmorillonite (chlorite and illite were dominant).

The clay mineral composition of sediments from northeastern Lake Michigan shows little variation with changes in depth of water, distance from source, or depth of burial. Due to the chemical content, pH and Eh of the lake water the three-layer clay minerals would not experience rapid diagenetic alteration in this environment. It can be concluded that the influence of source material is the primary control on the mineralogic character of the lake sediments.

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FIGS. 1–10, SEPTEMBER, 1961

OCCURRENCE AND ORIGIN OF CLAY PEBBLES ON THE EAST COAST OF JOHORE, MALAYA[1]

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ABSTRACT

Clay pebble occurences have been found along a stretch of about 10 km along the China Sea coast of northeastern Johore (Federation of Malaya).

They are derived from clay sheets, which alternate with sand layers in tidal flat deposits. The upper clay sheets lie between the high tide and low tide marks. At low tide the clay, if exposed, dries and forms cracks; wave action at the next high tide loosens flakes and rolls them while washing them ashore. Breakwater develops in a series of low rollers due to the flatness and shallowness of the sea floor. The clay pebbles are left behind at the high tide mark. Another deposit is found higher up the beach at a storm tide mark mainly as the result of monsoon storms.

The size of the clay pebbles ranges between 0.5 and over 30 cm, depending on the thickness of the clay sheet from which they are derived. Their roundness depends on the distance between the exposed clay sheet nip and the high tide mark. In this zone, a third migrating deposit of clay pebbles is usually present. After deposition, some pebbles may sink rapidly into the underlying wet sand, or sand may be blown over them. The clay is mainly provided by the rivers debouching in this area; the sand may be of marine origin, derived from rock capes in the neighbourhood.

INTRODUCTION

Clay pebbles have been described, among others, by Haas (1927) who reported about "clay balls," and by Gardner (1908) who found concretions formed by coagulation of clay during deposition, whereas Bell (1940) described armored mud balls. Shrock (1948, p. 331–334) discusses mud balls in pyroclastic deposits. Twenhofel (1950, p. 592) mentions "clay galls" which may form on some tidal flats from wet muds; a publication on clay galls has also been made by Burt (1930). The topic is discussed in general by Pettijohn (1957).

In the present paper the occurrence and the origin of clay pebble deposits on a coastal stretch in northeastern Johore (Malaya) will be discussed. The most characteristic part of this occurrence, around the outlet of a small river named Sungei[2] Takadai, was investigated in detail. In figure 1 the location of the area is depicted; full topographic information is given by the Topographic Map of Malaya 1:63,360, series HIND 1035, sheet 3H/7, Kuala Endau.

GEOMORPHOLOGY OF THE AREA

This part of the east coast of the Malay Peninsula is characterized by low hilly ridges, with few exceptions not exceeding 120 m in height, which may be part of a planation surface (feuchttropische Einebnungsflache) in the sense of Büdel (1957). They run in an approximately NNW–SSE direction, cutting the coast at an oblique angle. The ridges consist either of the Pahang Volcanic Series[3] or of the quartzite-shale series of Triassic age (Scrivenor, 1931). Intensive weathering has taken place in the upper 10 metres or more. Most of the area is under jungle.

The Endau river flows through this area in broad meanders (fig. 1), which have obviously shifted several times. Where the ridges abut on the shore, steep cliffs usually develop. The debris is deposited in bays and between the islands, so as to form a series of beaches, spits and tombolos. Swamps rather than lagoons, develop where a bar or tombolo has cut off a former part of the sea, for these parts of impeded drainage are rapidly filled up with clay. Furthermore vegetation is established easily on such tracts.

Most of the inland precipitation is carried to sea by a few main rivers. The young sediments of the coastal area however are drained by numerous small rivulets each

[1] Manuscript received July 7, 1960.
[2] Sungei = river.
[3] Mainly tuffs and rhyolitic rocks of Carboniferous-Triassic age.

with a limited inflow area. Both the main and the smaller rivers transport large quantities of clay towards the sea.

Many of the smaller rivers show 90° bends in their directions where they are forced to flow parallel to inland beach ridges. In the area under discussion, fossil beach ridges have been found as much as 5 km inland, directly south of the village of Endau.

In the proximity of the capes, pebbles and cobbles form the present beaches, but after a very short distance they rapidly decrease in size; the greater part of the beaches is sandy. Near the river outlets, however, the common picture is a tidal flat in front of and beside the actual outlet. These flats are built up of a sequence of sand and clay layers, the latter normally being 10–30 cm thick, although thicknesses up to 60 cm have been recorded. It is assumed that the sand is of marine origin and that the clay is provided by the rivers; field evidence is strongly in support of this assumption. The layers are sharply separated from each other; the clay layers are very heavy and homogeneous (fig. 10), whereas the sandy layers generally have variable clay contents.

Summarizing it may be said that the area has a tendency of outward coastal growth by deposition of sandy sediments which alternate with fluvial clays at the river outlets, and by the filling of swamps with clays. At the same time the exposed rock cliffs retreat landward, thus resulting in a progressive straightening of the coast.[4]

THE CLAY PEBBLE OCCURRENCES AROUND SUNGEI TAKADAI

The Sungei Takadai rises in the hilly zone at a height of only slightly above 30 metres. From its source it runs windingly mainly northward over about 5 km of ter-

[4] More detailed information on the geomorphology of this area may be found in a publication by the present author, which is to be submitted to the Journal of Tropical Geography.

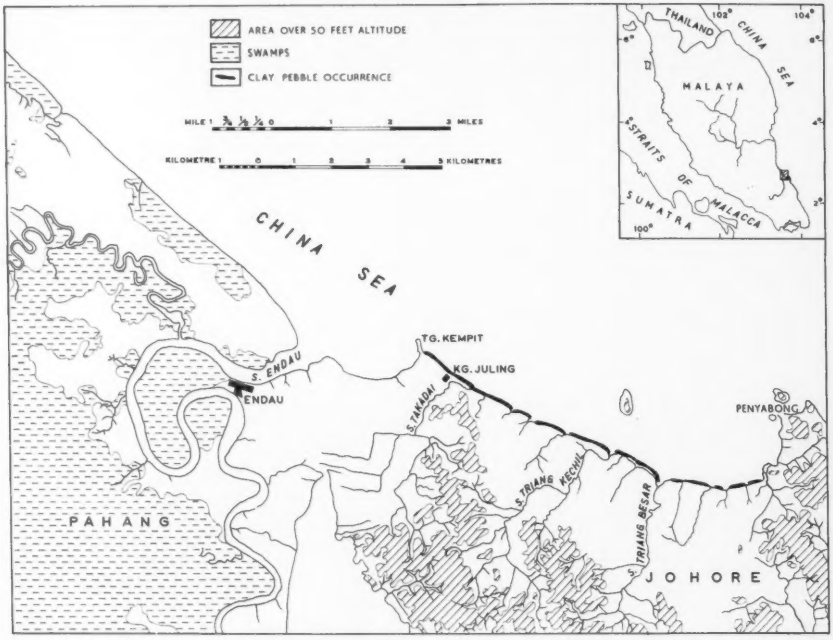

FIG. 1.—Location map.

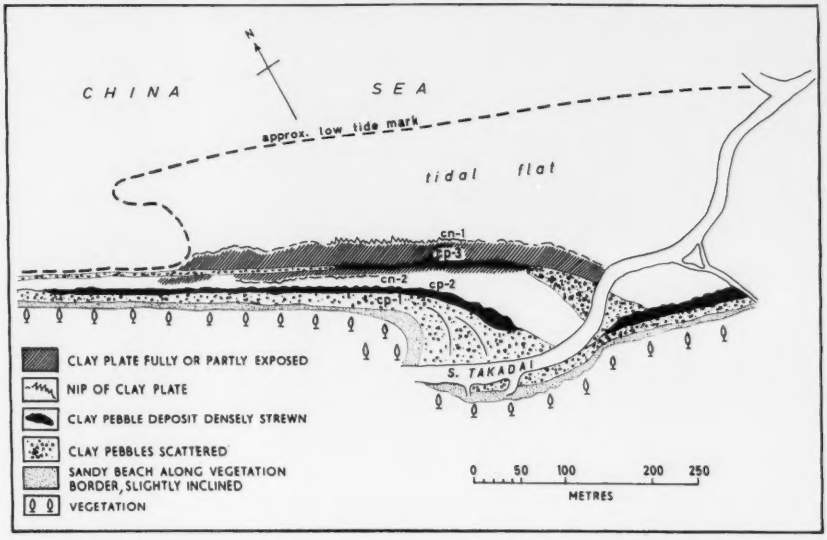

FIG. 2.—Clay pebble deposits northwest of Sungei Takadai outlet; early November 1959.

rain, receiving a number of tributaries. The heavy precipitation of 2.75–3.00 meters annually (Dale, 1959) gives this small river a considerable volume. Its course is influenced by the presence of inland beach ridges as described above. and has also been modified slightly in certain localities by drainage and irrigation works for rice fields. Its present outlet lies directly east of Endau; before reaching the sea the river is forced into a direction parallel to the coast by the most recent beach ridge.

At the river outlet the low tide mark fans out over quite some distance as a result of the increased sedimentation. A tidal flat is present here, consisting of a vertical sequence of sand and clay layers. Because of the tidal flat, a zone of very shallow water extends in front of the river and the adjacent beach at high tide. This has a very important effect in that it reduces the normal breakers to a gentle succession of very low rollers over the extremely shallow sea floor. The above features are characteristic of the whole stretch of the coast where clay pebbles occur. Wave action is most subdued during the period of the southwest monsoon. During the northeast monsoon, winds are stronger and correspondingly wave action is more powerful, but it is still conditioned by the tidal flat.

On both sides of the river outlet, clay pebble deposits were found; the situation as described in the following section applies to October–November, 1959.

The clay pebbles were found to have a greatest diameter ranging from 8 to over 30 cm (fig. 3). In figure 2 a sketch-map of the clay pebble occurrence northwest of the river outlet represents the situation in early November, 1959. This was just before the beginning of the northeast monsoon. Crossing the beach in the seaward direction, the following was observed. The vegetation ended at a sandy nip of about 1 meter high. This nip was bordered by a sandy stretch, sloping 2–4° seaward, with an average width of about 10 meters, except for the immediate vicinity of the river outlet, where a large sand plate extended. Next, a nearly horizontal part followed, sandy at the surface, again about 10 meters wide. On this part, scattered clay pebbles were present. In figure 2 this clay pebble occurrence is indicated as CP-1. The concentration of clay pebbles is obviously at a storm tide mark.

FIG. 3.—Selected clay pebbles. Upper left: length and cross section. Upper right: clay pebble beside its hole in which it is sinking into underlying sand. Centre: clay pebbles with shell fragments on surface; these were the only ones showing shells on surface, out of many hundreds.

The horizontal part of the beach was followed by an inclined sector with a seaward slope of 1½–2°. On this part, a second accumulation of clay pebbles was present. This one was much more densely strewn; densities of 30–50 or more clay pebbles per square meter were normal in this zone, which is indicated as CP-2 in figure 2, and also shown in figure 4. The deposit shows clearly visible cusps with an amplitude from apex to apex of 7–15 metres. Farther seaward there was a third zone of clay pebbles (CP-3, in fig.2) which was much less densely strewn than CP-2, and with less clear cusps, if any. The density of pebbles here varied from 4–20 per square metre, and showed a marked decrease farther from the river outlet.

These three zones extended for some 600 meters northwest of the river outlet. CP-1 and CP-2 were found to end abruptly, whereas CP-3 died out gradually. Northwest of this clay pebble area, the beach was all sandy and much more steeply inclined towards the high tide mark, namely 8° as against 1½–2° in the clay pebble zone. It was found that CP-2 lies just at the average high tide mark; CP-3 lies below high tide. Outside of the clay pebble occurrence, the first horizontal part of the beach and the nip at the vegetation border remained visible; the beach was very much narrower, though.

The clay pebbles are usually well rounded. Their structure is very homogeneous with-

FIG. 4.—CP-2, northwest of Takadai outlet; November 1959.

out any sign of zoning (fig. 3) which indicates that they do not undergo any outward growth. Their texture is heavy clay (fig. 10). Incidentally a shell fragment or such other items as a grain of gravel or a piece of rope were found in the clay pebbles. Nevertheless it is very clear that these items do not act as a centre for coagulation. In a following paragraph we shall deal with the origin of the clay pebbles. Southeast of the river outlet the picture was very much the same as northwest of the river.

FURTHER OCCURRENCE OF CLAY PEBBLES

Clay pebble occurrences very similar to the one depicted above, were found in the whole coastal stretch between Tg. Kempit at the outlet of the Endau River, and the Penyabong peninsula, that is over a distance of about 10 km (see fig.1) Their configuration always follows the same pattern: CP-1 at a storm tide mark higher up the beach; CP-2 as the main concentration just at the average high tide mark, and CP-3 less densely strewn, seaward from CP-2. It was ascertained that southeast of this area, as far as the outlet of the Jemaluang river,[5] no further clay pebble deposits occur. However in other places farther to the north along the Malayan east coast it is reported that clay pebble accumulations are to be found.

ORIGIN OF THE CLAY PEBBLES

In the river outlet areas and on the tidal flats, a vertical sequence of sand and clay layers occurs as described above. In the case of the S. Takadai the bedload of the river is obviously forced to settle on entrance into the sea although some clay may be carried alongshore for a short distance. The same happens to the other coastal rivers; since these rivers are usually at short distance from each other, the clay layers deposited in one outlet area may very well reach those of the next river.

Locally these clay layers are exposed. They may lie at the surface as clay sheets or plates, or their edges may be exposed in a nip, the rest of the clay being buried under a sand layer. At low tide the exposed clay is subject to drying and accompanying surface cracking (fig. 5). Towards the next high tide, the rising water comes with the previously described series of low breaking rollers (which seldom are more than 30–40 cm high), attacking the exposed clay along the cracks, forming a nip of maximally some tens of centimetres high (fig. 5). Thus, flakes of clay are loosened. Because of the shallowness of the sea floor, the area remains under breaking wave action as long as the water is over the exposure. This results in the clay flakes undergoing a continuous rolling and tossing, which gives them their rounded clay pebble shape. This rounding, of course, cannot be compared to that of solid rock pebbles since it is a matter of plastic modelling. With the rising water the flakes, now being converted into pebbles, are carried ashore and they finally are left at the high tide mark. This transfer is largely by traction and saltation.

The formation of cusps is the result of the pattern in which the water of any one wave reaches its outer mark on the beach. Consequently, the cusps are subject to changes. Under normal conditions, during quiet weather which prevails throughout the southwest monsoon and also during parts of the northeast monsoon, this process may account for the formation of CP-2. The CP-1 deposit higher up the beach is a similar result, but of storm or springtide conditions. On such an occasion it may be expected that CP-2 will be fully or partly destroyed, and that it will re-form after conditions have returned to normal.

CP-3 takes a somewhat different position. This is essentially a shifting deposit. As the high tide comes up, a clay exposure will be attacked. From that moment on, clay pebbles begin to form, leading finally to deposition at CP-2. As the water level starts falling again, the clay exposure will continue to be under attack with clay pebbles still being formed. As the water no longer reaches the high tide mark the pebbles are left at the outer mark of the water, and a thinly strewn zone of clay pebbles is thereby provided for on the seaward side of CP-2. Finally, the water level falls so low that the clay exposure is left dry and the supply of new clay pebbles stops. The cycle is repeated with the next high tide. The scattered clay pebbles of CP-3, together with those newly formed at the edge of the clay sheet, are now

[5] Some 45 km farther southeast along the coast.

FIG. 5.—Exposed clay sheet with cracks; November, 1959, low tide. Compare fig. 2.

washed ashore and deposited at the high tide mark, at CP-2. Then the water falls again, leaving a new CP-3 between the clay exposure and the high tide mark.

The situation in the Takadai area in early November 1959 was that one clay sheet was exposed both on its seaward and its landward side, giving rise to the formation of two nips, each of them under the attack of the breakwater. In figure 2, these nips are indicated as CN-1 and CN-2, respectively. In some parts, depending on the thickness of the sand cover, there was only one nip or none at all. The center of the clay sheet was at all places under sand, 5–30 cm in thickness. The position and thickness of such sand layers, however, may vary even from day to day. Below the exposed clay sheet, in borings another clay layer was found, and it may be presumed that more clay layers will be present deeper in the sequence. The Sungei Takadai was found to cut one of the lower clay sheets when cutting through the beach. It was found, moreover, that the northwestern and southeastern terminations of the clay pebble occurrences coincide with the termination of the clay exposure.

DAILY AND SEASONAL CHANGES

Daily changes.—On numerous occasions it could be observed how the configuration on the beach may vary within a few days. For example, southeast of the Takadai river outlet, the CP-2 deposit which continued from the one represented in figure 2, was well developed on October 27, 1959. Five days later the deposit was hardly visible. Many of the clay pebbles had sunk into the underlying sand, or they were covered with wind blown sand. Many clay pebbles too had been partly or fully destroyed in this five day period, though no strong winds or heavier seas were observed.

The sinking of clay pebbles may be favoured when the underlying sand is wet. Sunken pebbles stand a fair chance of being preserved in the sequence of layers since they are thus protected from direct water attack. The shape of the clay sheets also varies from day to day under the influence of the continuous breakwater attack. Contemporaneous sedimentation of clay adds to the changing picture on a longer time scale.

Seasonal changes.—The seasonal variations become clear after a comparison of figures 6 and 2. As stated previously, figure 2

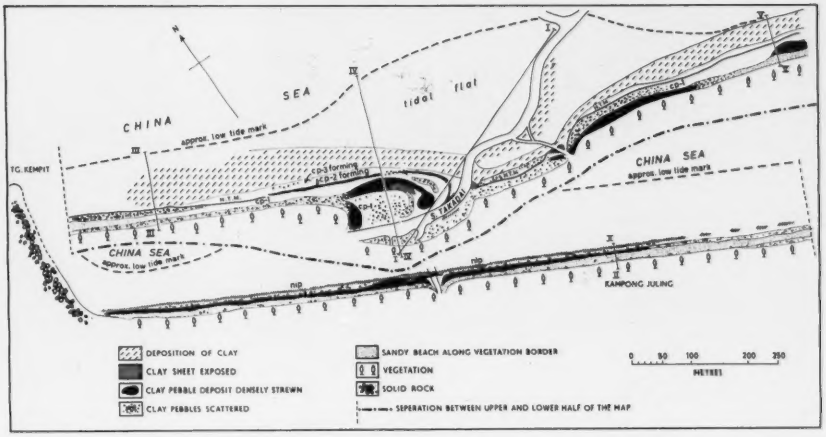

FIG. 6.—Clay pebble deposits in Takadai outlet area; April 1960.

represents the situation in early November, 1959. Figure 6 represents the situation as surveyed in detail in mid April, 1960. A section over the beach along the Takadai outlet is represented in figure 7. More sections are shown in figure 9. The map of figure 6 covers a greater area than the area shown in figure 2. The period in between the two dates of survey is the time in which most of the bad monsoon weather occurs, especially in December and January. In that period, stronger northeasterly winds and very intensive downpours prevail. No measurement of the height of the waves during this period can be given, but it is certain that during storm periods, they are considerably higher than under "normal" conditions. Large masses of wreckage, uprooted trees, etc. at the storm tide mark in April illustrated the rougher conditions during the northeast monsoon months.

For the Takadai outlet area, the changes from November–April can be summarized as follows.

On the sand bar along the river outlet, CP-1 and CP-2 have been preserved; differences in configuration will be noted on comparison of the two maps. The clay pebbles of these deposits showed surfacial signs of weathering: cracking and fracturing. The deeper ones had been preserved better. New sand had been deposited upon the bar; perhaps some new clay pebbles were supplied, too. This could not be ascertained.

In the northwestern direction, CP-1 was found to have shifted and to lie now directly along the vegetation border (fig. 6). These pebbles also showed signs of surface attack. This change of situation must be a direct result of the monsoon storms, with the water coming high up the beach. In the quiet season, the storm mark shifts seaward, giving rise to the situation as depicted in figure 2, whereas the clay pebbles higher on the

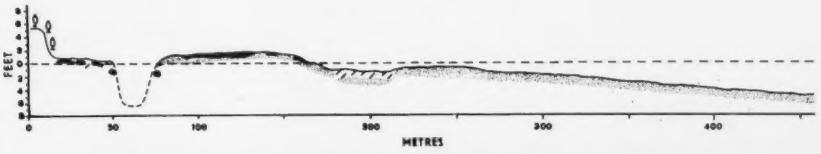

FIG. 7.—Section I showing surface sediments. For location and legend see fig. 6.

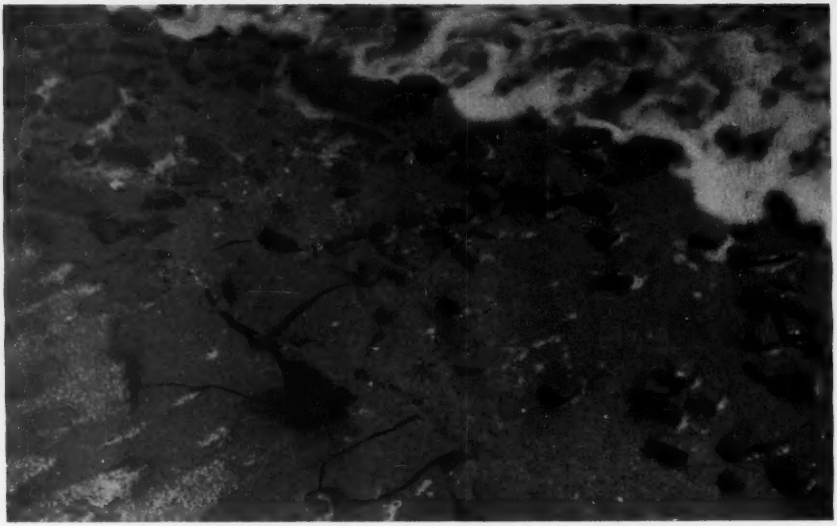

FIG. 8.—Attack of clay sheet in Kampong Juling area; April 1960; about half an hour before high tide.

beach are buried under wind blown sand; an occurrence which becomes evident already early in the SW monsoon season.

CP-2, in this area, is subjected to even greater changes. On April 12, 1960, CP-2 was observed to have been reduced to almost nil. During the monsoon the clay sheet from which it was derived has been eroded fully, and moreover the original deposit as seen in figure 4, has been destroyed during the monsoon. Only a few pebbles, buried deeply in the sand, have been preserved. At this time (April, 1960), it was found that a new clay sheet was in the early stage of deposition (fig. 6), approximately on the same area as where the hardened clay sheet was found in November 1959. Two days later, it was found that the clay sheet had already begun to harden a little. Moreover, at falling water, the ebb streams cut small gullies through the clay. Together with the first stage of cracking, this provided sufficient opportunity for the waves to loosen miniature flakes and roll them to miniature clay pebbles, which may be referred to as clay grains. These grains were again concentrated at the high tide mark as CP-2, whereas the shifting CP-3 deposit was also found seaward of CP-2. Nevertheless, the deposition of the clay sheet still goes on at high tide. As the consolidated clay sheet grows thicker, the size of these clay grains will increase and finally a situation similar to the one depicted in figure 2 may be expected.

In November 1959 it was found that northwest from the termination of the Takadai clay pebbles, the beach became sandy and remained free from any clay pebbles for a considerable distance. During the northeast monsoon, a deeper lying clay sheet has become exposed so as to form an extensive clay pebble deposit in the area around Kampong Juling. This clay pebble area now reaches almost as far southward as the northwestern end of the Takadai area pebbles (fig. 6). In figure 8 the attack on this clay sheet and the clay pebbles thus formed, are shown. The configuration of this Kampong Juling deposit shows the same features as are normally found. CP-1 lies along the vegetation border, thinly strewn at the surface with many pebbles already

buried under sand. CP-2 again is the main concentration, situated at the high tide mark. Densities of 30–50 clay pebbles per square meter, in the center of the zone, are quite common. These pebbles in general are not as round as those in the Takadai area, since the clay exposure is close to the high tide mark in a nip about 2 feet high (compare fig. 9, section II), so that the distance of transportation and the period of submergence are short. Yet a distinctive CP-3 zone shifts between the exposure of the clay sheet and the high tide mark (fig. 8).

The above is an illustration of the manifold and rapid changes that occur on these clay pebble deposits. It may be assumed that similar changes take place throughout the coastal stretch in which clay pebbles occur.

PHYSICAL PROPERTIES OF THE CLAY AND SAND

Various properties of the sand and clay of this clay pebble area have been analyzed. The results are listed in this paragraph. It would be premature, however, to draw conclusions from these analyses, since no data are available yet for the properties of the rocks and younger coastal sediments in the northeastern part of Johore. And it is these rocks and sediments which may, to a smaller or greater extent, influence the properties of the sediments in this clay pebble location, since they act as a source region for the fluvial clay, and also for the marine sand which is obviously derived from rock capes under wave attack.

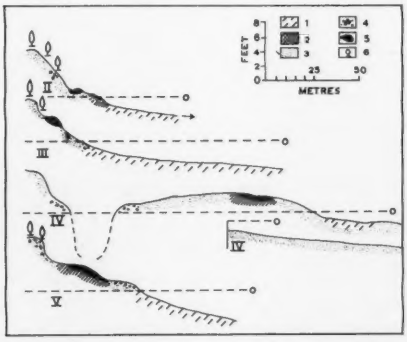

FIG. 9.—Sections across the beach showing surface sediments. For locations see fig. 6. Legend: 1—deposition of clay; 2—clay sheet exposed; 3—sandy deposits; 4—clay pebbles scattered; 5—clay pebble deposit densely strewn; 6—vegetation.

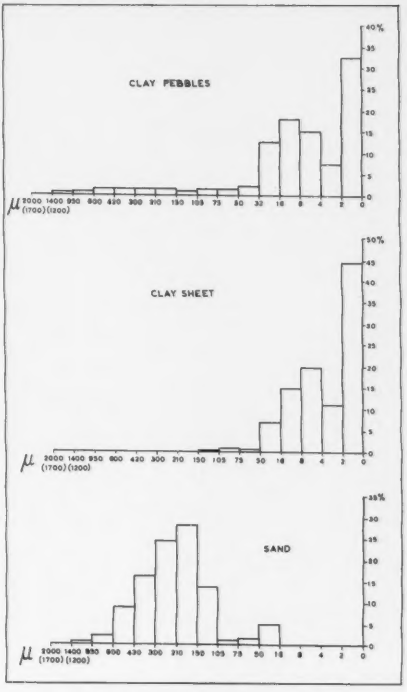

FIG. 10.—Grain size distributed histograms.

Granulometric analysis of the clay sheet, the clay pebbles and the sand higher up the beach was carried out in the Physical Geographical Laboratory of the University of Amsterdam. The results are represented in figure 10. The similarity in grain size distribution of the clay sheet and the clay pebbles once more illustrates their relationship. The heavy clayey texture is obvious, 90 and 73 percent of the particles being smaller than 16 microns. It is interesting to note a striking double peak, one in the clay fraction smaller than 2 microns, the other in the 16–8 or 8–4 micron interval. Consequently the q-values are low,

$$\left(q=\frac{\text{percentage of particles smaller than 2 microns}}{\text{percentage of particles smaller than 16 microns}}\times 100\right)$$

namely 49 and 45. The high percentage in the 16–2 micron interval may be attributed to the settling of these particles on the contact of the fresh riverwater with the salt seawater, which produces locally brackish conditions.

The sand of the higher beach is a rather well-sorted fine to medium sand; to the sorting both marine transport and wind blowing have contributed.

Heavy mineral analyses were carried out by the author in the University of Malaya, Singapore. The different sources of the sand and the clay are reflected in the different mineral compositions (table 1). Analysis of these source areas is being undertaken now.

The clay mineral composition of the clays is summarized in table 2, after analysis by Mr. Th. Levelt (Physical Geographical Laboratory, Amsterdam, Holland). As analysis of the weathering products of the hinterland is still in progress, the above analysis is given here without further commentary. The kaolinite was found to have a low to moderate degree of crystallinity.

TABLE 1.—*Heavy mineral analyses*

	Clay Sheet Mutual %	Sand Mutual %
Tourmaline	6	26
Zircon	59	37
Rutile	4	2
Anatase	—	4
Brookite	—	2
Garnet	5	1
Cassiterite	10	7
Cyanite	—	1
Andalusite	4	13
Sillimanite	1	—
Staurolite	11	6
Epidote	—	1

TABLE 2.—*Clay mineral analyses (Analyst, Th. Levelt)*

	Clay Sheet	Clay Pebbles
Montmorillonite	—	—
Illite	30	25
Kaolinite	65	70
Hydrargyllite	tr.	tr.
Quartz	5	5

CONCLUSIONS

The occurrence of clay pebbles in the area surveyed is governed by the following conditions. Clay sheets must be exposed at least at their edges at low tide so as to form cracks. The slope of the sea bottom, seaward from the high tide mark, must be gentle so as to cause the breaking waves to develop a very extensive zone of low rollers. A thickness of the clay sheet between 10 and 30 cm seems most favourable for clay pebble formation; hard sand layers facilitate the rolling of the flakes loosened from the edges of the clay sheet into pebbles.

Three main zones of clay pebbles can be distinguished. One lies at a storm tide mark and a second one, usually the main concentration, at the high tide mark. A third zone of pebbles is constantly migrating between the clay exposure and the high tide mark.

When deposited on wet sand, some of the pebbles may sink rapidly into it, thus being protected from weathering or wave attack. Such pebbles stand a fair chance to remain preserved in the sequence of layers.

Climatic influences seem not to be decisive. However, intense sun radiation largely favours cracking while the clay is exposed. The heavy monsoon rains may account for extra supply of clay by the rivers, which is deposited in and around the outlet area, so as to provide new clay sheets.

It will be clear that tidal flat conditions, especially in a humid tropical climate, are an almost ideal site for such clay pebble formations. It may be of interest to investigate whether the occurrence of clay pebbles in consolidated sedimentary rocks is related to rocks deposited under tidal flat conditions.

ACKNOWLEDGMENTS

The author wishes to convey his gratitude to the following persons who cooperated in this investigation. Prof. Dr. J. P.

Bakker and Dr. H. J. Müller (Amsterdam) enabled the grain size analyses to be performed. Mr. Th. Levelt (Amsterdam) carried out the clay mineral analyses. Prof. Kiang Ai Kim (Singapore) allowed the author to carry out analyses in the Chemistry Department, University of Malaya in Singapore. Prof. W. A. R. Wikkramatileke (Singapore) read and corrected the English of the manuscript. Last but not least, the author had most useful field assistance from his wife, Mrs. W. Nossin.

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REMANENT MAGNETISM IN SILTS OF PLEISTOCENE AGE FROM NORTH AMERICA[1]

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ABSTRACT

Remanent magnetizations of 64 silt specimens of Kansan, Illinoian, Sangamon, and Wisconsin age are reported. All, except some of the Illinoian specimens, are apparently normally magnetized. The primary magnetic fabrics of the Kansan, Sangamon, and Wisconsin specimens may have been deformed slightly by the consolidating effects of overriding Pleistocene ice sheets. Twenty-two specimens from the basal part of an Illinoian glaciolacustrine silt display a confusion of north-magnetic-pole plots, 21 of the plots lying between latitudes 18°S to 88°N and longitudes 7° to 176°W. This apparent confusion of the magnetic field corresponds to the time of the Illinoian glacial maximum in southern Indiana. Nine specimens from the upper part of the Illinoian silt unit, representative of post glacial-maximum time, are normally magnetized.

INTRODUCTION

Irving's (1959, 1960) comprehensive survey and analysis of paleomagnetic pole positions indicates that most of the Pleistocene positions that have been determined are assignable to either early or very-late Pleistocene time. A few pole positions are known for "middle-Pleistocene" time, but the precise moments within the epoch to which the pole positions may be assigned are unknown. In this report are presented the results of an investigation into the remanent magnetizations of silt samples of middle to late Pleistocene age from well-understood stratigraphic sections of North America.

In 1958 and 1959, Harrison collected oriented samples for magnetic analysis from Indiana and Illinois; Terasmae and P. F. Karrow collected samples from Ontario and Quebec. Magnetic analysis was performed by J. Mead, using a "spinner"-type magnetometer, while specimen preparation and data calculations were accomplished by Harrison. The authors are indebted to Dr. Mead for making the magnetic measurements.

LOCATIONS, STRATIGRAPHY, AND SEDIMENTATION

Sixty-four specimens, representing small time intervals of the Kansan, Illinoian, Sangamon, and Wisconsin stages of the Pleistocene epoch were studied. Magnetic pole positions (table 1) were calculated by averaging the magnetic vectors of each group of specimens and plotting the mean vector according to the methods described by Graham (1955). Where the dispersion of the magnetic inclinations for a group of closely associated specimens was such that the semi-axis, d_X, of the elliptical error area around the pole (at a probability of 95 percent) was less than 14, the vectors were not averaged; instead, the individual pole positions, as determined from the magnetic vectors for each specimen, were listed individually (table 1, col. 6).

Kansan Samples

The oldest of the Kansan samples were cut from a bed of silty alluvium in western Illinois. Horberg (1956) considered the alluvium to be early Kansan in age. The bed of noncalcareous, unoxidized silty clay exhibited occasional layers of well-preserved twigs of conifers; it cropped out 6 in above stream level near 39° 52′ 30″ N. Lat., 91° 18′ W. Long. Because it was overridden by the Kansan and Illinoian ice sheets, the bed is not normally consolidated but overconsolidated.

Samples of a proglacial silt, considered by Wayne (1958, p. 10) to be of Kansan age, were taken in west-central Indiana (near 39° 28′ 30″ N. Lat., 86° 54′ 30″ W. Long.). The samples were cut from the laminated lake

[1] Manuscript received August 11, 1960.

TABLE 1.—*Magnetic-north-pole positions obtained from remanent magnetizations of specimens of silts of Pleistocene age from North America*

Footnotes are to studies that give stratigraphy and (or) sedimentation at sampling localities.

Stratigraphic Position (1)	No. Loc. (2)	No. Spec. (3)	Pole Positions Lat. (4)	Pole Positions Long. (4)	Vertical Sampling Profile (5) Spec. No.	(6) Pole Positions Lat.	(6) Pole Positions Long.
Wisconsin:							
Glacial maximum[1]	1	5	56N	138E			
Scarborough beds[2]	2	11	73N	35E			
Sangamon:							
Upper Don beds[3]	1	2	69N	175E	22	18S	86W
					21	12N	148W
Middle Don beds[3]	1	2			20	20N	145W
					19	8S	74W
Basal Don beds[3]	1	5	84N	170W	18	1N	176W
					17	48N	96W
Illinoian:					16	4N	152W
					15	50N	145W
Post glacial maximum[4]	1	9	71N	33W	14	4N	7W
					13	88N	156W
Glacial maximum[5]	2	22	31N	112W	12	35N	81W
					11	80N	88W
					10	0	77W
Kansan:					9	80N	150E
					8	24N	71W
Preglacial maximum[6]	2	6	64N	57E	7	61N	141W
					6	4N	84W
Early alluvium[7]	1	2	80N	65E	5	45N	156W
					4	20N	42W
					3	36N	107W
					2	46N	41W
					1	31N	173W

[1] Harrison, 1958, table 1, no. 6, and appendix, no. 6; also, Harrison, in preparation.
[2] Terasmae, unpublished studies; (specimens from basal 1.0 meter).
[3] Coleman, 1914, 1933.
[4] Wayne, 1958, p. 12, units 7 and 8; (3 specimens taken along one bed are in agreement).
[5] Wayne, 1958, p. 12, unit 4, (basal 0.6 meter).
[6] Wayne, 1958, p. 10, unit 3; (3 specimens taken along one bed are in agreement).
[7] Horberg, 1956, appendix, no. 13; (change location to NW¼ SE¼ of section).

silt described by Wayne (1958, unit 3); this bed is overconsolidated, owing to the stresses of overriding Kansan and Illinoian ice sheets.

Illinoian Samples

Several samples of a laminated glaciolacustrine silt of Illinoian age were taken near 39° 16′ 30″ N. Lat. and 86° 25′ W. Long. in south-central Indiana. Petrography of the silts is given by Smith (1956) and the stratigraphy by Wayne (1958, p. 12). Specimen cubes cut from samples from the middle zone of the silts were too weakly magnetic to permit accurate measurements, and so the magnetic data of table 1 are presented in two groups, representing the upper and lower samples. The valley in which the lake silts were deposited was dammed upstream and downstream by southward moving ice of Illinoian age. Assuming Flint's (1955) rates of overland ice advance and retreat, the lake may have existed 2500 to 4500 years while the ice advanced to and retreated from the Illinoian drift boundary. It is possible, therefore, that the lowermost

Illinoian silts were deposited about the time of the glacial maximum (table 1, col. 1), but that the uppermost silts were deposited during glacial retreat, some 2000–4000 years later. The uppermost samples of the lake silts are overlain by only 3 ft or so of laminated deposits and then by 6 ft of outwash sands and gravels that attest, probably, to *rapid* melting of the retreating Illinoian ice. Finally, 17 ft of noncalcareous silt, that is probably mostly loess, overlies the outwash sands and extends to the upland surface.

All of the cubes for magnetic analysis were cut from the more clayey layers of the laminated beds, where the beds were calcareous, unoxidized, and undesiccated. Where sampled, the beds were composed of about 30 to 40 percent clay (mainly illite and chlorite) and 55 to 65 percent very-fine silt. Smith (1956, nos. 18–27) reported that a series of channel samples through the silts had an average *p*H of 8.0 and contained detrital magnetite and organic material in the $>49\mu$ fraction (amounting to less than about 2 percent of his samples). The greatest dips observed in the undisturbed, laminated beds were 3° to the northeast. The silts were undoubtedly in a state of normal consolidation and were sampled in the undisturbed state.

There are many similarities between the glaciolacustrine silts of Illinoian age of this study and the glaciolacustrine silts of Wisconsin age from New England that were studied by Johnson, Murphy, and Torreson (1948). Both the Illinoian and the Wisconsin silts are laminated, physically and chemically undisturbed, and normally consolidated. The magnetic vectors found in the specimens of the basal Illinoian silts (table 1), however, are much more steeply inclined than are the vectors that were found by Johnson and others in the Wisconsin silt specimens. The nine specimens from near the top of the Illinoian lake clays, on the other hand, have inclinations (table 1) comparable to those found in the Wisconsin silt specimens from New England.

Sangamon Samples

Silt beds of Sangamon age (table 1) were sampled at three places around the periphery of the Don Valley Brickyard pit near Toronto, Ontario (43° 40′ N. Lat., 79° 22′ W. Long.). Coleman's studies (1914, 1933) of the macroscopic flora and fauna, and Terasmae's study (1958) of the pollen spectra from the Don Beds indicate that a portion of these beds was deposited when the mean annual temperature was probably 3–5° C warmer than the present at Toronto. The beds are an alternating series of alluvial and lacustrine sands, silts, and gravels. They have been consolidated to an unknown degree as a result of stress induced by the thick sheet of overriding Wisconsin ice. Cubes for magnetic analysis were cut from samples of calcareous, unoxidized silt beds.

Wisconsin Samples

The oldest samples of Wisconsin age (table 1) were obtained from the Scarborough beds at the Don Valley Brickyard pit at Toronto, Ontario. The Scarborough beds, where sampled, were lacustrine silts and were calcareous, unoxidized, and undisturbed. They were consolidated by the overriding Wisconsin ice sheet to the same extent, probably, as the underlying Don beds of Sangamon age.

An interstadial silt unit of possible early-Wisconsin age (Harrison, in preparation), dated at >38,000 B.P. (W-814, W-578), was sampled near 39° 52′ 30″ N. Lat. and 86° 18′ W. Long. Where sampled, the silt beds were calcareous, unoxidized, horizontally laminated, and undisturbed. The silt unit has been consolidated (Harrison, 1958, tab. 1, no. 6) to a probable value of 43.0 kg/cm² (to a minimum value of 26.0 kg/cm²); this amounts to an *overconsolidation* of, probably, 38.4 kg/cm², the present effective overburden pressure being 4.6 kg/cm².

REMANENT MAGNETISM

The remanent magnetism ranged in intensity from approximately 2×10^{-5} cgs unit per cc in some of the specimens of Illinoian age to 45×10^{-5} cgs unit per cc in some specimens from the Don beds. A *primary* magnetic fabric undoubtedly was imparted to the silts at the time of their formation, when detrital ferromagnetic particles became oriented in the earth's magnetic field as they settled out of aqueous suspension.

POSSIBLE MODIFICATIONS OF THE PRIMARY MAGNETIC FABRIC

Irving (1959) has considered the possibilities whereby sediments fail to preserve the direction of the geomagnetic field imparted to them at the time of their formation. He indicates that little is known of the effects upon the primary magnetic vector produced by post-depositional overconsolidation of detrital sediments. All but the Illinoian sediments of this study are overconsolidated to varying degrees; their in-place void ratios are many times less than those which would be expected if they had been consolidated only by the sediments that presently overlie them.

Consider the silt unit deposited about the time of the early Wisconsin glacial maximum (table 1). The north-magnetic-pole position (table 1, col. 4) obtained for the silt unit lies a few degrees south of the limit of the expectable secular variation (approximately the 60th parallel) and in the *eastern* hemisphere. Because the bed has been overconsolidated by about 38.4 kg/cm², one might expect that the average inclination of the primary magnetic fabric has been "flattened" toward the horizontal plane during sediment consolidation by overriding Wisconsin ice some 425 m (1400 ft) thick. One can infer that silt rebound (Harrison, 1958, no. 6), since overconsolidation by now-vanished ice, has not been of sufficient magnitude to permit "relaxation" to the degree required for restoration of the primary magnetic fabric. It is probable that all of the other specimens (except the Illinoian) have been overconsolidated to an even greater extent than the "Wisconsin maximum" samples. Because of this fact, there is some difficulty in relating the magnetic pole positions found for these samples to "absolute" geographic pole positions.

The following information may be cited in support of this view. Four silt specimens were obtained from a bed of normally-consolidated Pleistocene lake silts in western New Hampshire (43° 41′ 30″ N. Lat., 72° 16′ 49″ W. Long.). The bed occurs in an exposure of "varved clays" of the same type and age as those studied by Johnson and others (1948). One specimen from the bed was stressed in the soil-mechanics laboratory at Dartmouth College to a value of 160 kg/cm². This is equivalent to roughly 1,775 m (5825 ft) of overriding glacial ice. A specimen cube the size of one of the three unstressed specimens was then made up by combining two thicknesses of the stressed specimen in identical horizontal orientation. The results indicated that the magnetic vector of the specimen stressed in the laboratory had been flattened about 10° toward the horizontal when compared with the vectors of the unstressed specimens.

PALEOMAGNETISM OF ILLINOIAN SPECIMENS

The normally-consolidated specimens of Illinoian silts (table 1) are thought likely to exhibit primary magnetic fabrics. Detailed studies by Johnson and others (1948), of very similar sediments of Wisconsin age indicate that laminated, normally-consolidated, glaciolacustrine silts are among the best of the clastic sediments to measure for purposes of paleomagnetic reconstructions. There is at present no reason to suspect that the magnetic vectors of the Illinoian specimens are not representative of primary magnetic fabrics.

Assuming that the geomagnetic field throughout geologic time tends to be symmetrical about the earth's axis of rotation, one is tempted to infer shift of the geographic north pole from the paleomagnetic data of these Illinoian specimens. The wide dispersion of the magnetic pole positions (table 1, col. 6) for the 22 specimens must be satisfactorily explained, however, before shift of the geographic pole can be postulated. The wide dispersion might possibly reflect the *disturbance* of the geomagnetic field during a slight shift of the earth's axis of rotation.

Momose's (1958) measurements of Japanese volcanic rocks and lacustrine sediments indicate continuous shift of the centered dipole from Lat. 75° N. to Lat. 70° S. during Pliocene times. Momose's rate of magnetic polar shift is considerably less than that implied by our two groups of Illinioan specimens (table 1, cols. 1, 4). Theoretically, however, the rate of pole migration could be much greater (Gold, 1955, p. 528). The effect of rapid geographic polar shift on

the geomagnetic field, unfortunately, is indeterminable.

SUMMARY AND CONCLUSION

Measurements of the remanent magnetizations of silt specimens of Kansan, Sangamon, and early Wisconsin age indicate magnetic-north-pole positions within the expectable secular variation (north of the 60th parallel, approximately), but these pole positions may not be the true positions during the time of silt deposition. A secondary magnetic fabric may have been imparted to the silts as a result of the stresses exerted upon them by overriding Pleistocene ice masses followed by only partial intergranular readjustment after unloading by the ice.

Some evidence is given for movement of the centered geomagnetic dipole during the time of the Illinoian glacial maximum in Indiana. This evidence is contrary to other data (Irving 1959, 1960) that indicate considerable stability of the geomagnetic field throughout the Pleistocene epoch.

It is felt that additional measurements of the remanent magnetization of clastic sediments of Pleistocene glacial and interglacial age will prove profitable. Wherever possible, the sampled sediments should be in a state of normal consolidation and should occur in geologic sections that are well-understood stratigraphically.

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FIGS. 1–3, SEPTEMBER, 1961

CYCLIC BEDDING IN PURE CARBONATE ROCKS FROM SOUTH-CENTRAL FRANCE[1]

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ABSTRACT

Field and laboratory analyses of Oligocene carbonate rocks from the Massif Central region of France emphasize the dolomitic nature of the rocks and their cyclic bedding. The section displays an alternation between a very finely-crystalline, white calcite-dolomite mixture and a similar lithology, richer in dolomite and mottled with irregular tan dolomite masses. Thickness of cycles averages 0.5 meters.

INTRODUCTION

This paper reports on a cyclically-bedded sequence of limestone and dolomite of Oligocene age in the western portion of the Massif Central in France. The outcrop described is located near Arpajon (Cantal) in the Aurillac Basin (fig. 1), one of several basins in the Massif Central in which Oligocene sediments have been deposited and preserved.

Field observations were made during the summer of 1957 and laboratory analyses carried out during the following winter in the Department of Geology, Princeton University, Princeton, N. J.

Thirteen cycles were recognized within the 9.7 meters of section measured and sampled in detail. Two facts emerge from the X-ray diffraction analyses of the rock types found in the cycles: (1) each cycle is made up entirely of calcite and dolomite; and (2) the ratio of dolomite to calcite increases greatly upward through each cycle.

According to Jung (1946, p. 86), deposition of Oligocene sediments began in Sannoisian time in the southern part of the Massif Central. Deposition occurred in grabens formed by early Tertiary (Pyrenean) faulting. The basal sediments rest on the smooth surface of the "Pénéplaine Eogène," a well-developed erosion surface formed during the late Mesozoic and early Cenozoic when the area of the Massif Central was emergent. The Sannoisian sedimentation was confined to the southern parts of the Massif Central, but the succeeding Stampian deposits attain great thickness and wide distribution in basins throughout the crystalline massif. The Stampian section contains at its base some arkosic red beds with minor amounts of lacustrine limestone. Throughout much of Stampian time the subsiding troughs received great quantities of fresh and brackish water limestones and marls so that along the western edge of the largest of these basins, the Limagne, the Stampian reaches 1750 meters in thickness. The border faults of these basins were the site of the beginnings of vulcanism in the massif which, during the Miocene and Pliocene, found expression in the construction of the immense

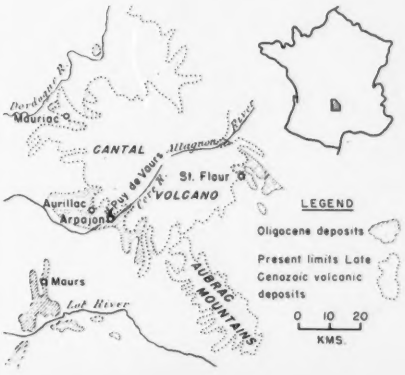

FIG. 1.—Index map of Puy de Vaurs locality.

[1] Manuscript received September 2, 1960.
[2] Present address, Brooklyn College, Brooklyn, N. Y.

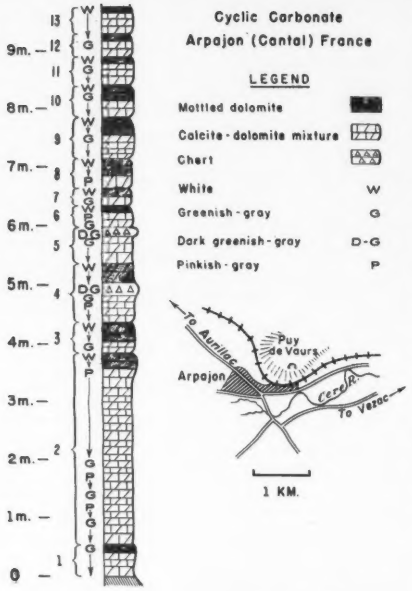

FIG. 2.—Columnar section, Puy de Vaurs excavation, Arpajon (Cantal) France.

Cantal and Mont Dore volcanoes. Volcanic activity has continued into the Pleistocene epoch. Quaternary erosion has scored the surfaces of the volcanoes and in part excavated the Oligocene basins.

LITHOLOGIC DESCRIPTIONS

The section discussed here is located a few score meters north of the railroad station at Arpajon (Cantal) halfway up the south side of a hill, the Puy de Vaurs. The section is well exposed in an abandoned lime excavation. The face of the workings (fig. 2) displays white, thin-bedded carbonate rocks overlain by darker, thin-bedded nodular carbonate rocks that top the exposure. The carbonate section rests upon red shales and sandstones exposed in the slopes below the excavation. Jung (1946, p. 122) has dated the basal white carbonates, which concern us in this paper, as Middle Stampian and the overlying carbonates as Upper Stampian and Aquitanian in age. The shale upon which the carbonate section rests is presumed by Jung to be Lower Stampian in age.

The 9.7 m of section exposed between the talus at the base of the excavation and the nodular carbonate rocks demonstrates a striking alternation between two lithologic types. One is a thin-bedded, chalky-appearing carbonate rock of soft hues, either light greenish-gray or, less commonly, light pinkish gray. Thin sections of this type show it to be very finely crystalline, and mineral determinations could not be made from them. Field tests for the presence of dolomite were positive, and X-ray diffraction tests, carried on later in the laboratory, showed the lithology to be a mixture of calcite and dolomite (fig. 3). The other lithologic type is similar to the first except for its whiter color and that it contains tan, irregularly-shaped dolomite masses that become more prominent towards the top of the bed. X-ray tests on the tan mottled mass of this unit (fig. 3) indicate the presence of dolomite almost to the exclusion of calcite. X-ray analyses of

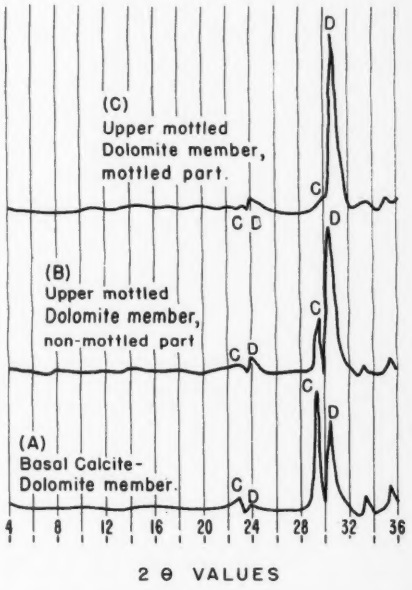

FIG. 3. X-ray diffractometer traces of a Puy de Vaurs cycle using nickel filtered copper radiation. C—calcite, D—dolomite.

the non-mottled part of this unit demonstrate a combination of calcite and dolomite intermediate between that of the mottled dolomite masses and that of the non-mottled rock type described previously. In all carbonate samples on which X-ray tests were run there was no indication of any components other than calcite and dolomite.

Thirteen cycles can be recognized in the Puy de Vaurs section. The basal member in each is the non-mottled unit while the mottled dolomite is the upper member. The contact between cycles is sharp while that between members of a cycle is less sharp and somewhat gradational. All contacts are parallel throughout the face of the excavation. Interrupting the orderly succession of cycles are two thin chert beds that occur midway in the measured section. Variations in the cyclic succession can be noted in cycle No. 2 (fig. 2), where differences in color within an abnormally thick cycle may indicate it to be a composite of several cycles, and in cycle No. 5 where the upper member is not developed. Cycles No. 5 and No. 6 may represent one irregular cycle. The average thickness of the cycles, excluding the unusual cycle No. 2, is 0.51 meters.

COMMENTS

Several comments can be made concerning the significance of this outcrop. First of all, previous descriptions of these sediments have emphasized the presence of argillaceous limestones (marls), a description not borne out by this study. Furthermore, to the best of the author's knowledge, the presence of dolomite in this section has not been recorded. It would be well to re-examine the other Oligocene exposures of the Massif Central in the light of the findings of this paper.

It is noteworthy that the paleogeographic conditions that produced the arkosic red bed deposition of the Sannoisian and early Stampian stages changed rapidly so that middle Stampian deposition is marked by pure calcite and dolomite precipitation without introduction into the basin of deposition of detrital quartz and clay.

Alderman and Skinner (1957) recently reported the occurrence of contemporary primary dolomite in southeastern Australia. These sediments are finely crystalline white calcite or calcite and dolomite mixtures and appear to be similar to the rocks described in this paper. Alderman and Skinner state that the dolomitic sedimentation is largely controlled by the elevated pH of the water which, in turn, is caused by the high rate of plant growth. It is not unreasonable to propose that similar conditions prevailed when the Oligocene sediments of the Aurillac Basin were laid down.

The thin annual cycles of the Australian dolomite occurrence do not appear to hold the answer for the origin of the Puy de Vaurs cycles. The latter are too thick to be considered annual. Furthermore we do not know what percentage of the dolomite in the Puy de Vaurs cycles is primary and what percentage is caused by post-depositional alteration of the sediments. Perhaps cyclic weather changes, controlling the pH through the intermediary of plant life, may be the principal control of the deposition of this cyclic section.

ACKNOWLEDGMENTS

This study forms a by-product of a program of joint archeologic-geologic studies in the drainage basin of the Dordogne River under the direction of Dr. Hallam L. Movious, Jr., Peabody Museum, Harvard University, Cambridge, Massachusetts. Professor Sheldon Judson of Princeton University, Princeton, New Jersey was in charge of the geological studies. Financial support was provided entirely by grants from the National Science Foundation.

Dr. Judson and Dr. Jerome Regnier of Mobil Oil Company de Venezuela critically read the manuscript. Mobil Oil Company de Venezuela assisted in the completion of this paper and Sr. Eduardo Troconis T. of that company kindly drafted the figures. The author wishes to thank these people for their invaluable assistance.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 31, NO. 3, PP. 456–466
FIGS. 1–5, SEPTEMBER, 1961

PETROGRAPHY OF THE TEAYS-MAHOMET VALLEY DEPOSITS[1]

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ABSTRACT

Samples of sediments from just above the bedrock of the valley bottom of the ancient Teays and Mahomet Rivers were taken in unglaciated regions of West Virginia and Ohio, and from beneath glacial drift in Ohio and Illinois and analyzed for heavy mineral content. Analysis of the West Virginia samples showed a mature suite of minerals characterized by rounded grains of tourmaline and zircon, some epidote, a high muscovite content, and the absence of pyroxenes. The Ohio samples contained abundant angular grains of hornblende, a few pyroxene grains, as well as much angular zircon and tourmaline. Analysis of the Illinois samples showed an abundant garnet content with much hornblende, epidote, a small amount of pyroxene, and in general displayed mineralogy typical of the Pleistocene tills. The mature mineralogy of the sediments in the unglaciated area of West Virginia indicates derivation of these sediments from the Paleozoics of the folded Appalachians rather than from the Piedmont. Mineralogy of the sediments in the valley bottom beneath the glacial drift suggests that the Teays-Mahomet alluvium has been scoured from these locations and has been replaced by glacial drift.

INTRODUCTION

Purpose

For several years the state geological surveys of Ohio, Indiana, and Illinois, and the U. S. Geological Survey have been investigating ground water resources in regions of buried bedrock valleys now covered, for the most part, by thick accumulations of glacial drift. As additional data from water, gas, and oil drilling operations and from seismic investigations have been received, more accurate location of these bedrock valleys and the direction and amount of their gradients has been established. The manner in which the late Tertiary drainage pattern was affected by the advance and retreat of the various Pleistocene glaciers is still the object of some controversy. These bedrock valleys and the sediments within their confines represent a long, multi-cycled time of erosion, extending from the Tertiary through the Pleistocene. An examination of the heavy mineral suites from the alluvium of the ancient Teays and Mahomet River Valleys was undertaken for the purpose of determining the mineralogy, degree of weathering, and provenance of the alluvium.

Area of Study

The Teays River flowed from its headwaters region in the Piedmont of Virginia and North Carolina (according to Stout and Lamb, 1938) northwestward to Chillicothe, Ohio (fig. 1). From this point, the valley has been traced beneath the glacial drift in a northwestward direction to the central Ohio-Indiana border, and thence westward through Indiana and Illinois. In western Illinois, the valley turns southward to join the bedrock valley of the present Illinois River (Horberg, 1945), and then continues southward to the ancient Mississippi River. This preglacial stream system, approximately 950 miles in length, served as a major drainage feature in the Appalachian Province, and through the Central Lowlands Province. In the latter physiographic region, all Paleozoic systems except the Cambrian were exposed (Fenneman, 1938) at the time the Teays-Mahomet River eroded its channel.

Previous Investigations

G. F. Wright (1890) traced the channel of the "Teaze" River from the Kanawha River below Charleston, West Virginia, to the Ohio near Huntington, interpreting it as a former course of the upper Kanawha. W. G. Tight, in 1903, described the physiographic history of this eroded stream channel in West Virginia, and recognized three erosion periods with intervening times of "slack water" conditions when silt was deposited.

[1] Manuscript received September 12, 1960.

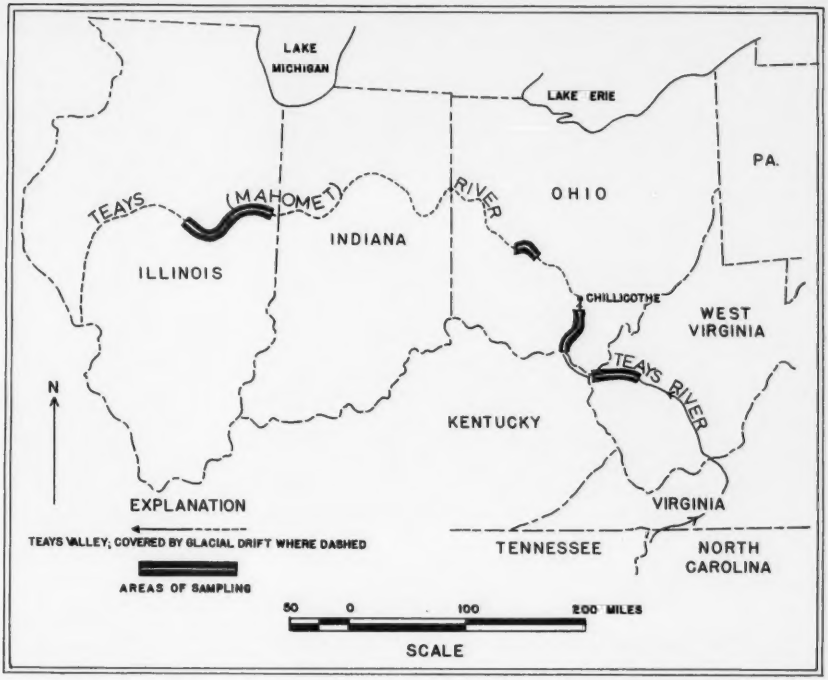

FIG. 1.—Course of the preglacial Teays-Mahomet River and areas of sampling (modified after Norris and Spicer, 1958).

Stout and Schaaf (1931) described the Minford silt, which occurs in the Teays channel where it becomes obscured by glacial drift south of Chillicothe, Ohio, as a laminated deposit 10 to 80 feet thick which was deposited in ponded streams during prolonged flood conditions. These silts were shown to contain more than 50 percent of well preserved sericitic mica (illite). Aside from rutile, chlorite, and some "free silica," there appears to be a paucity of accessory minerals in these deposits. A blue-gray, fine textured deposit of siliceous silt immediately below the Minford silt was referred to as the alluvium of the preglacial streams. The basal sands and sandy silts of the valley were reported to be of fine to medium texture, colored yellow with iron oxide coatings, and containing "a few percent of mica and clay matter. . . . ." Stout and Lamb (1938) stated the possibility that the sericitic mica (illite) in the Minford silt was derived from the schists of the Piedmont, thus requiring extension of the headwaters of the Teays well into this area. The location of this headwaters area was reaffirmed by Stout, Ver Steeg, and Lamb (1943), who reported an average stream gradient of 10 inches per mile.

Lockwood (1954) examined macroscopically and microscopically hand auger samples of Teays sediments in three Ohio counties and reported:

"Heavy mineral content is low, the suite composed of zircon, garnet, muscovite, tourmaline, hornblende, limonite, epidote, andalusite, leucoxene, ilmenite and hypersthene. With the exception of zircon, all of the minerals are angular in shape. Scattered throughout the base of the sand are Sharon type pebbles, more angular sandstone, and chert pebbles and boulders."

Foster and Buhle (1951) and Horberg (1953) discussed the nature and age of glacial drift deposits in Illinois and their possibilities as aquifers, as have many others (for example, Rosenshein, 1958, and Norris, 1959), for various locations along the presumed course of the Teays-Mahomet River. Apparently, however, statistical work on heavy minerals in the fill of these ancient bedrock valleys is lacking.

PROCEDURE

Sampling

In the region of Huntington, West Virginia, where Teays sediments crop out, specimens of basal sand and lacustrine silts and clays were taken by Professor Raymond E. Janssen of Marshall College.

Seven hand auger samples of basal Teays sand (Lockwood, 1954) were obtained from locations along the channel at depths down to 47 feet. Size data of these specimens are shown in table 1.

Two samples sent by Mr. Stanley E. Norris were taken from the same well at depths of 220–230 feet, and 230–235 feet respectively. These samples from test hole number 7 (Norris and Spicer, 1958) were located in Teays sand just above bedrock on the northeast side of the main Teays channel.

All Illinois samples were chosen from the files of the Illinois Geological Survey on the basis of their location with respect to the Mahomet channel as shown on contours of the bedrock topography (Horberg, 1950). Wherever possible, samples from wells that penetrated to the deepest part of the bedrock valley were selected, and analyses of the sands and silts immediately above the bedrock surface were performed. Areal distribution throughout the state, as well as location of well sites relative to the stream channel, governed the choice of samples. Figure 2 shows the locations of these Illinois samples. Locations of all samples are given in appendix A.

Processing of Samples

All samples used were oven dried at 90° C for 48 hours prior to sieving, and then separated through a 9 mesh Tyler sieve to remove grains over 1.98 mm in size. Where grain size indicated a prospective poor yield of heavy minerals within the desired size fraction (such as the clay and fine silt specimens from West Virginia), the entire sample received was used. Two hundred grams of Teays sand was used from each location where the original quantity permitted.

Each sample was wet sieved through 60 mesh (.250 mm openings) and 325 mesh Tyler screens. The 325 mesh screen, with openings of .043 mm, was chosen to insure a sufficient yield of heavy minerals in the size range in which these minerals are normally more abundant.

The desired size fraction was then heated in dilute hydrochloric acid to remove soluble carbonates and iron. After effervescence ceased, each sample was washed and oven dried. Bromoform was employed as the heavy liquid in the separations, and the heavy minerals were mounted in Canada balsam. Two hundred grains per slide were identified with the aid of a petrographic microscope, provided at least 100 clear grains were included in the count. In samples having a high percentage of

TABLE 1.—*Size data of seven Ohio samples collected by William Lockwood*

Sample number	Depth	Weight whole sample	Weight .250–.043 mm fraction	Weight heavy fraction	Percent* heavy minerals of fraction
124	22–42 ft	116.4 g	46.9 g	.098 g	.21
129A	14–18	157.5	13.7	.057	.42
133	18–27.5	135.4	35.7	.170	.48
141	14–18	116.9	30.4	.137	.45
142	12–14	108.6	12.5	.045	.36
145	34–41	113.9	51.8	.570	1.10
146	34–47	127.2	39.7	.153	.39

* Slide rule computations.

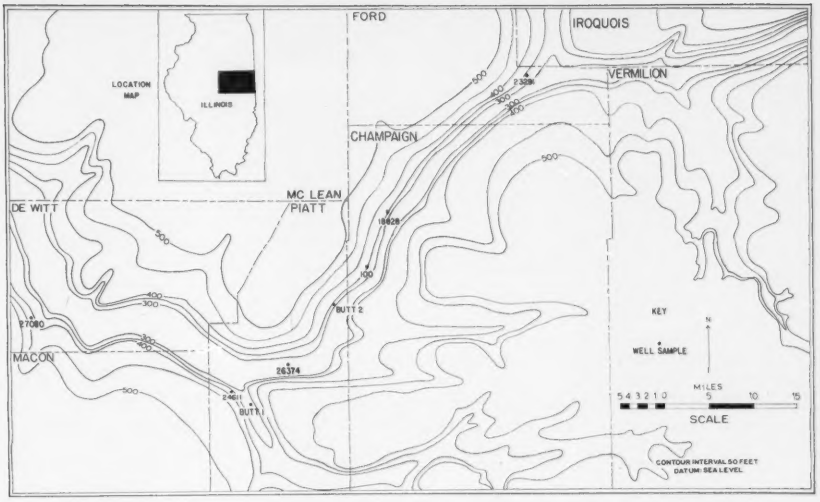

FIG. 2.—Samples localities in relation to bedrock topography of east-central Illinois (after Horberg, 1950). Contours above 500 feet are omitted.

TABLE 2.—*Mineral species by percent*

	Sample	Andalusite	Aug.-Diop.	Biotite	Enstatite	Epidote	Garnet	Hornblende	Hypersthene	Kyanite	Muscovite	Olivine	Rutile	Staurolite	Titanite	Topaz	Tourmaline	Trem.-Actin.	Zircon	Black Opaques*	Other Opaques*	Unknown	Clear Grains	Total Grains
West Va.	A			2		30		5			40			1			5		17	22	30		100	208
	B-1			3		2		6			17						5	1	66	14	74		100	203
	B-2			2		12	1	6		1	12						11		55	11	42		100	214
	C					2	3	5			12						10		66	9	94	2	100	214
	6150					3		1			88						3	1	4	8	8		170	200
	6155										tr.								tr.	tr.	tr.			
	6158										tr.						tr.		tr.	tr.	tr.			
	7392										100												200	200
	7662					tr.	tr.	tr.									tr.		tr.					
	7797										100												200	200
	7807						tr.	tr.			tr.								tr.					
Ohio	124	1				9		2			69						2		17	5	29		133	200
	129A		1	6		13	3	24		1	2		1	1			4		42	5	40	2	111	200
	133			4	4	16		9	1		6			1	1	6	19		30	13	44	3	100	229
	141				3	26	3	20	1	1			1	5	2	1	10		26	16	43	1	100	246
	142			5		29	4	17			1		1		3		9		31	16	38		100	217
	145	1		3	2	12	2	55			6						3		16	11	24		130	200
	146			6		1	6	7			38			2			18	1	20	4	58	1	100	238
	7a		2		2	3	22	55	2			3					1	5	4	15	8	1	155	200
	7b	1	1		2	11	19	40	3		4	1		1	1		6	2	8	24	14		126	200
Ill.	23291		1		3	18	14	59	2								1		2	8	22		141	200
	18828					8	20	64	1						1		2		3	9	39	1	104	200
	100				1	16	34	40			2			1	2		1		3	1	64		100	285
	Butt 2				1	20	8	62	3								2	2		8	26	2	132	200
	26374		1		1	24	17	50	3								1		2	12	8	1	160	200
	Butt 1					25	15	56	1								1		2	8	57		100	200
	24611		2		2	13	20	51	3		1				1		3	2	2	16	20		129	200
	27080		2		3	7	12	49	6			18					1			12	9	2	158	200

* Percent of total grains.

opaques, when 100 clear grains were not recorded in the first 200 grains, additional grains were counted to bring their sum total up to 100.

Evaluation of Mineralogy

Table 2 shows the percent of mineral species at the various localities. Four samples of the basal Teays Valley sand in West Virginia, samples A, B-1, B-2, and C (whose locations are listed in appendix A) yielded sufficient heavy minerals for analysis. The exceedingly high abundance of such minerals as zircon and muscovite, and the relatively high abundance of tourmaline in comparison to hornblende and epidote, constitute a suite of stable minerals and attest to an environment that has favored destruction of the more unstable minerals. Olivine and the pyroxenes are absent and hornblende accounts for no more than six percent of the clear heavy grains of any sample. Discounting the possibility that the scarcity of hornblende in these deposits is entirely due to weathering in the outcrop area, this mineral, if the suite was derived from the headwaters region of the Teays in the Piedmont, may not have survived to the region of sampling. The wide size range of the samples studied (fig. 3) would normally compensate for a dearth of the species within one or two size fractions.

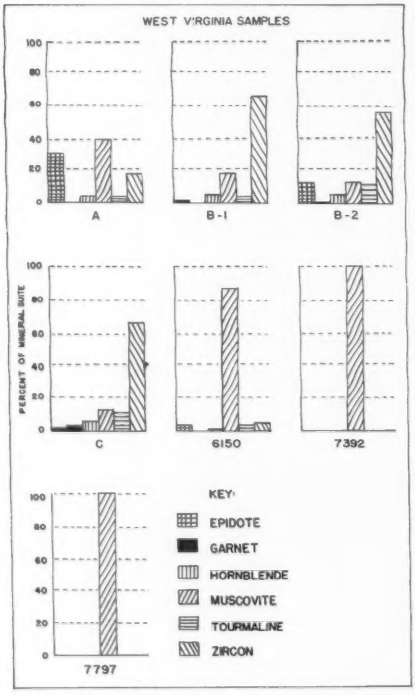

FIG. 3.—Histograms of chief mineral occurrences in West Virginia deposits.

The meager occurrence of garnet in these samples is also noteworthy in the light of conflicting reports regarding the stability of this mineral. Allen (1948), studying the effects of heavy mineral weathering in Illinois glacial deposits, concluded that garnet was consistently more stable than hornblende in place. Pettijohn (1957, p. 506) similarly rated the persistence of garnet, and classified it rather close to tourmaline in durability. Dryden and Dryden (1946), examining minerals from deposits in the Maryland-Pennsylvania region, found garnet to weather more rapidly on the outcrop than hornblende, and if substantial amounts of garnet survived the weathering of the parent rock, this mineral may not have persisted to any notable extent after deposition. However, complete destruction of the species during transport seems unlikely.

With regard to the origin of the pebbles and larger clastic fragments associated with these samples, Professor R. E. Janssen (personal communication) wrote that many of these coarser clastics were probably derived from the Piedmont or Blue Ridge regions, and Cross and Schemel (1956) wrote:

"From the vicinity of the Kanawha River southwestward to the limits of the Ohio Valley in West Virginia, some boulders gravel and sand derived from igneous and metamorphic rocks are found in deposits laid down in the drainage system of the ancient Teays River. These may have come in part from eastern West Virginia and western Virginia, but some almost certainly came from far out on the Piedmont of North Carolina and Virginia, the headwater area of that former stream. An excellent place to collect these is in road cuts west of West Virginia Route 17 at Scary Station . . ."

Samples 7392 and 7797 were composed almost entirely of clay and fine silt, and barely yielded enough material larger than 43 microns for analysis. The other clay and fine silt samples produced only a trace of heavy minerals. These samples, taken from a stratigraphically higher position than the Teays sand, yielded a heavy mineral fraction of 100 percent muscovite. The mineral and textural composition of these specimens, and their organge, buff, or brown color, a result of oxidation, suggest an origin and environment of deposition similar to the Minford silt described by Stout and Schaaf (1931).

Analysis of those samples obtained by Mr. William Lockwood from the unglaciated portions of the Teays Valley in southern Ohio reveals a consistently high precentage of zircon. Many of the zircon grains display a greater degree of angularity than those identified in the samples from West Virginia. A sharp increase in hornblende (to a maximum of 55 percent) is accompanied by the appearance of the rather unstable pyroxenes. It is of interest to note that the mineral suite of the samples from West Virginia persists to the segment of the Teays immediately south of the glacial drift in Ohio, although the fill of the bedrock valley in this segment shows an undoubted mineralogical association with the till deposits further north. Thus, the important appearance of orthorhombic and monoclinic pyroxenes takes place while the percentages of zircon and tourmaline, and, in two cases, muscovite remain high.

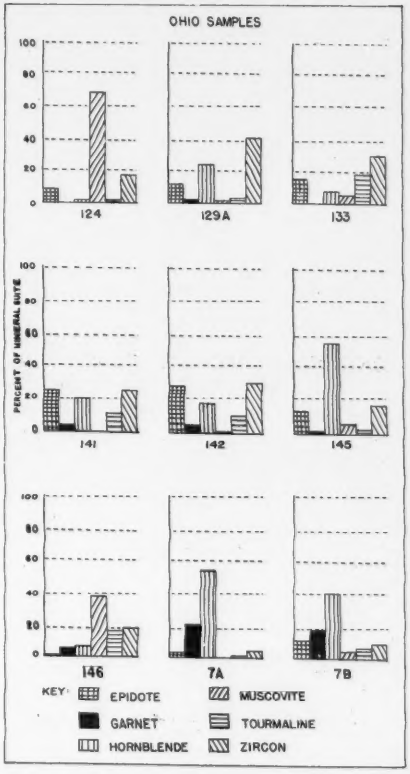

FIG. 4.—Histograms of chief mineral occurrences in Ohio deposits.

Macroscopically the samples of test hole 7, acquired by Mr. S. E. Norris from the Teays Valley in Madison County, Ohio, displayed the pebble shapes and poor sorting characteristic of till (fig. 4). The abundance of coarse silt and sand in these specimens produced a comparatively substantial yield of heavy minerals. The zircon, tourmaline, and muscovite that accounted for such a large proportion of the samples obtained from locations further upstream formed only a small percentage of these samples. Aside from a few well rounded grains of tourmaline and zircon, the angular, fresh appearance of the heavy mineral suite was evident.

The mineralogy of the well samples obtained from the Mahomet channel in Illinois bear no apparent relation to their easternmost counterparts. Hornblende, garnet, and epidote are particularly abundant, while zircon, tourmaline, and muscovite are scarce. None of the latter three minerals accounts for more than three percent of the clear heavy grains in any one sample. The presence of pyroxenes, and the general angularity of the grains further indicate a lack of relationship to those sediments which fill the Teays channel of West Virginia. They do indicate, however, on the basis of heavy mineral studies of approximately 350 Pleistocene till samples made by the writer for the Illinois Geological Survey, a direct relationship to those tills which now overlie

TABLE 3.—*Sankoty and Tertiary sandy-gravel analyses*

Sample	Garnet	Zircon	Hornblende	Epidote	Staurolite	Kyanite	Tourmaline	Muscovite	Hypersthene
P-368	8		50	22	2				6
P-73		26	7	6	27	7	19	7	
P-82		48	2		14	5	17	9	

P-368.—Sankoty sand. Borrow pit, S. side of valley. NW¼ SE¼ SW¼ Sec. 18, T. 15 N., R. 10 E., Bureau Co., Hennepin Quad., Illinois.

P-73.—Tertiary sandy gravel. Roadcut: west side of road, west of Commence, Missouri. SE¼ NW¼ NE¼ Sec. 26, T. 29 N., R. 14 E., Scott County, Mo.

P-82.—Tertiary sandy gravel. Roadcut: Intersection in Richfield. NE cor. NW¼ Sec. 17, T. 35 N., R. 6 W. Adams Co., Illinois.

(These samples were kindly supplied by Dr. H. B. Willman.)

the Mahomet Valley in Illinois. Sample 27080, from a well drilled in the south side of the valley in western DeWitt County, is unique because if contains olivine, which accounts for 18 percent of the clear heavy grains identified. The unusually high amount of this comparatively rare detrital mineral is best explained by a release from a basic or ultrabasic boulder close to the well. Figure 5 shows by histograms the chief mineral occurrences in the Illinois deposits.

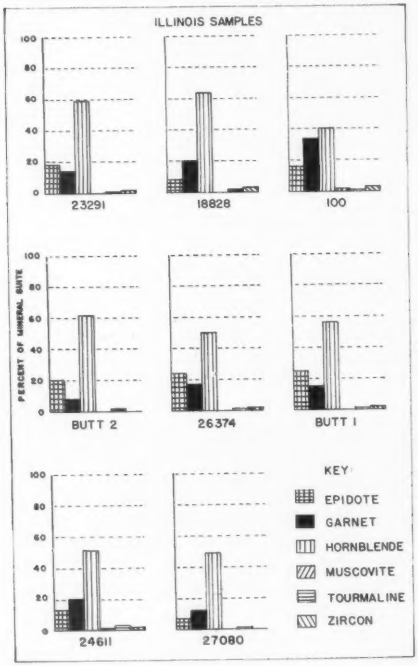

FIG. 5.—Histograms of chief mineral occurrences in Illinois deposits.

In a further attempt to evaluate the mineralogy of the basal Mahomet sediments, one sample of Sankoty sand, defined as " . . . the lowermost sand and gravel deposit which underlies glacial till or related outwash and overlies bedrock along the Ancient Mississippi Valley" (Horberg, 1953), and two samples of Tertiary Lafayette-type sandy gravels were analyzed (table 3). With regard to the origin of the Tertiary Lafayette gravels and associated clastics, Potter and Pryor (1960) wrote: "Although occasional grains of kyanite, sillimanite, and staurolite have been noted in some of the Paleozoic sandstones of the upper Mississippi Valley, the abundance of these minerals in Cretaceous and Tertiary sands shows that the mica schists and gneisses of the Piedmont region were the major source."

A close mineralogical similarity of the Sankoty deposit to that of the Mahomet channel was noted, while analyses of the Tertiary Lafayette-type gravel showed the mineralogy to be quite different from the Mahomet specimens, particularly in the percentages of staurolite, kyanite, tourmaline, and zircon. To further cite Horberg, although the composition and texture of the Sankoty sand (based chiefly upon light mineral and macroscopic observations) denotes a source area different from that of the Mahomet sand, its association with glacial drift is unquestioned.

It is possible that the well samples, taken

TABLE 4.—*Weathering ratios**

Sample	Fraction Zi+To/Hb	Ratio	Sample	Fraction Zi+To/Hb	Ratio
A	22/5	4.41	146	38/7	5.44
B-1	71/6	11.82	7a	5/55	.09
B-2	66/6	11.00	7b	14/60	.35
C	76/5	15.20	23291	3/59	.05
6150	7/1	7.00	18828	5/64	.08
124	19/2	9.50	100	4/40	.10
129A	46/24	1.92	Butt 2	2/62	.03
133	46/9	5.45	26374	3/50	.06
141	36/20	1.80	Butt 1	3/56	.05
142	40/17	2.35	24611	5/51	.10
145	19/55	.35	27080	1/49	.02

* Slide rule computations.
Key: Zi = zircon; To = tourmaline; Hb = hornblende.

from various locations on the cross section of this ancient bedrock valley system, are not contemporaneous with each other. Wayne (1956), and Thornbury (1958) wrote of the strath terraces along the Teays-Mahomet of Indiana and Illinois, the so-called Deep Stage, which represents the last erosion cycle in the formation of these valleys, and the controversy over when these levels were cut (particularly the Deep Stage). In selecting sample locations from the channel beneath the glacial drift, specimens of the fill from the deepest portion of the valley were chosen wherever possible, in the hope of obtaining material representing the true river alluvium.

In West Virginia, the absence of a metamorphic suite of heavy minerals in a sediment containing many igenous and metamorphic pebbles and boulders is unusual. It is also unusual (according to Thornbury, 1958, and in the opinion of the writer) that a stream whose headwaters region is in the Piedmont Province should flow in a westerly direction for nearly 1000 miles, to drain ultimately into the Gulf Embayment. The genesis of this major preglacial stream system remains as one of the more interesting problems of geomorphology.

Weathering Ratios

While ratios are best employed to illustrate a weathering variation in vertical sections of soil profiles, the use of the method here shows the marked differences in mineralogy at several locations along the bedrock channel. Zircon, tourmaline, and hornblende were chosen because these minerals are usually present in every mineral count, zircon and tourmaline being relatively very stable, while hornblende is relatively unstable. Thus, the ratio of the sum of the two more stable minerals to the hornblende is used as an indicator of the maturity which the entire specimen has attained. Six West Virginia clay-fine silt samples failed to yield satisfactory amounts of heavy minerals, and are excluded.

The high zircon+tourmaline/hornblende ratios obtained for the samples of Teays sand in West Virginia (table 4) show the comparative scarcity of hornblende relative to zircon and tourmaline; this high value continues to a lesser degree through most of the Teays samples from Ohio. In Illinois, the extremely low ratios reveal a different type of mineralogy: tourmaline and zircon occur in relatively very small amounts while hornblende is quite abundant. In effect, these ratios reveal a less weathered suite of minerals in the downstream segments of the Teays-Mahomet Valley than in the headwaters area.

CONCLUSIONS

Presence of abundant zircon, tourmaline, and muscovite in the West Virginia samples suggests a derivation of these samples from the sedimentary rocks of the folded Appalachians, rather than from the metamorphic rocks that underlie much of the more easterly Piedmont Province. Significantly, the Teays sand appears to be mineralogically mature in the unglaciated areas of West

Virginia in contrast to its immaturity in the areas covered by glacial drift in Ohio and Illinois. The entire suite of heavy minerals in the unglaciated area indicates an exclusive origin in the Paleozoic formations; a virtually complete lack of characteristic metamorphic minerals in the 13 samples analyzed from West Virginia seems to support this. Mineral persistence data (Dryden and Dryden, 1946; Pettijohn, 1957) preclude the possibility of the chemical destruction of these minerals through the distance from the region of the Piedmont to the sites from which the samples were taken, despite the fact that the Teays was a mature valley with a low gradient. Factors of mechanical breakdown during transport, or complete postdepositional destruction of the metamorphic suite would not seem to explain satisfactorily the absence of these minerals in the samples studied. Thus, there is nothing in the heavy mineral suites obtained from the unglaciated locations of West Virginia to indicate origin from metamorphic rocks.

The presence of a high percentage of the relatively unstable minerals such as hornblende and the inclusion of the pyroxenes in the mineral counts of the Ohio and Illinois samples, represent a marked change in the mineralogical maturity of the valley sediments. If the stream valley sediments are from "normal" upstream sources, the expected sequence is one showing increased mineralogical maturity westward, or downstream from the source area. The Ohio and Illinois samples analyzed in this study are not part of such a sequence; therefore they indicate a definite origin in glacial deposits.

The theory of a glacial drift origin for the sediments now occupying the Mahomet Valley of Illinois at the locations sampled is further supported by a mineralogical similarity between these specimens and a representative sample of Sankoty sand, and by a dissimilarity with the mineral suite of the Tertiary Lafayette-type sandy gravel.

The well samples taken from the Teays-Mahomet Valley in glaciated regions, although in close proximity to the bedrock bottom of the valley, resemble the glacial deposits of the various Pleistocene stages. This implies at least for these locations, that any alluvium existing in the channel prior to the Pleistocene was scoured away, and that subsequently this valley network was filled by glacial drift and eventually buried under thick accumulations of the glacial material.

ACKNOWLEDGMENTS

I wish to express appreciation to Professor Jack L. Hough for his assistance and advice in this work.

I am also indebted to Dr. H. B. Willman for making available well samples from the files of the Illinois Geological Survey, from which all the Illinois specimens were taken, and for providing laboratory facilities at the Illinois Geological Survey.

Special appreciation is due Professor Raymond E. Janssen of Marshall College, for supplying the samples from West Virginia. Mr. Russell A. Brant, of the Ohio Geological Survey, and Mr. Stanley E. Norris of the U. S. Geological Survey, Columbus, Ohio, provided samples from the various locations in Ohio. Mr. John P. Kempton of the Illinois Geological Survey encouraged me in the investigation.

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APPENDIX A

LOCATIONS OF SAMPLES

West Virginia

St. Albans Quadrangle

B-1 State Route 34, near Poplar Creek: Lat. 38° 27′ 30″ N., Long. 81° 56′ W. Elevation 680 feet.
B-2 Same as B-1, except about 10 feet lower on contact of bedrock. This is the Morgantown Sandstone, Conemaugh age, which formed the bedrock under the Teays River here.
C Teays Valley Road opposite church near Scary: Lat. 38° 26′ 30″ N., Long. 81° 51′ 30″ W., Elevation +700 feet. This is at the uppermost end (hanging end) of the abandoned section of the Teays Valley where it begins to descend to the Kanawha River.
6150 Teays Valley Road and Scary Creek Road: Lat. 38° 27′ N., Long. 81° 51′ 30″ W., Elevation +700 feet.
6155 State Route 34, near Poplar Creek: Lat. 38° 27′ N., Long. 81° 56′ W., Elevation +700 feet.
6158 Across road from Mt. Vernon School along State Route 34: Lat. 38° 27′ N., Long. 81° 57′ W Elevation 710 feet.
7797 Lat. 38° 22′ 30″ N., Long. 81° 49′ W., Elevation 760 feet.

Milton Quadrangle

A Northeastern edge of Hurricane, State Route 34: Lat. 38° 26′ N., Long. 82° 01′ W., Elevation 680 feet.
7662 One-half mile NE of Milton: Lat. 38° 26′ 30″ N., Long. 82° 00′ 30″ W., Elevation +700 feet.
7807 NW edge of Culloden: Lat. 38° 25′ 15″ N., Long. 82° 04′ W., Elevation +700 feet.

Guyandot Quadrangle

7392 Pea Ridge Road and Nickle Plate Drive: Lat. 38° 24′ N., Long. 82° 23′ W., Elevation +600 feet.

Ohio

Sciotoville Quadrangle

141 Lat. 38° 58′ 10″ N., Long. 82° 48′ 7″ W.
142 Lat. 38° 57′ 50″ N., Long. 82° 48′ 7″ W.
145 Lat. 38° 58′ 40″ N., Long. 82° 46′ 30″ W.
146 Lat. 38° 57′ 20″ N., Long. 82° 48′ 7″ W.

Waverly Quadrangle

124 Lat. 39° 02′ 20″ N., Long. 82° 52′ 35″ W.
133 Lat. 39° 04′ 42″ N., Long. 82° 56′ 20″ W.
129A Lat. 39° 00′ 42″ N., Long. 82° 52′ 40″ W.

Illinois

Ford County

23291 Center of west line, Sec. 8, T. 23 N., R. 10 E.

Champaign County

18828 330′ N., 330′ E., NW¼ SE¼ NW¼ Sec. 23, T. 21 N., R. 7 E.
100 NW¼ NW¼ Sec. 21, T. 20 N., R. 7 E.

Piatt County

Butt 2 NW¼ NE¼ NW¼ Sec. 12, T. 19 N., R. 6 E.
26374 Near Wilson and Front St. Sec. 12, T. 18 N., R. 5 E.
Butt 1 NE¼ SE¼ SW¼ Sec. 17, T. 17 N., R. 5 E.
24611 36′ N., 34′ W. of SW c. Sec. 25, T. 18 N., R. 4 E.

DeWitt County

27080 2100′ from S line, 250′ from E line Sec. 16, T. 19 N., R. 1 E.

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FIGS. 1–2, SEPTEMBER, 1961

RELATIONSHIPS BETWEEN POROSITY, MEDIAN SIZE, AND SORTING COEFFICIENTS OF SYNTHETIC SANDS[1]

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ABSTRACT

The relationship between porosity, median size, and sorting coefficients has been investigated by studying synthetic sands with lognormal size distributions and various median sizes and sorting coefficients. Poorly sorted sands are considerably less porous than well sorted ones, and porosities show an inverse linear relationship to sorting coefficient except in very well sorted samples in which porosity increases more rapidly than sorting. Porosity is independent of median size in well sorted sands but appears to decrease slightly with increasing median size in poorly sorted sands. The relationship between median size and porosity probably results from the fact that larger grains have a higher sphericity and tend to pack more closely together than smaller, more irregularly shaped grains.

INTRODUCTION

The relationship between porosity and texture of sandstones, though qualitatively well known, has received comparatively little quantitative study. The object of the present paper, therefore, is to describe some quantitative relationships between porosity, median size, and sorting in synthetic sands.

Previous work has shown that porosities of uniformly sized spheres should vary from 47.6 percent if the spheres have a cubic packing to 25.9 percent, if the spheres are rhombohedrally packed (Graton and Fraser, 1935). Highest porosities are invariably found in assemblages of particles which are all of the same size (perfectly sorted), and porosities decrease as the range of particle sizes increases (Fraser, 1935). Theoretically, porosity is independent of grain size (Slichter, 1899), although in natural sediments porosity tends to increase with decrease in median grain size (Fraser, 1935; Griffiths, 1952). This relationship, which is apparently contrary to theory, may be explained by change in shape of natural particles with grain size. In general, smaller grains are less spherical and tend to pack together less regularly than the more spherical larger grains, consequently forming a greater porosity in finer grained sediments.

[1] Manuscript received July 8, 1960.
[2] Currently employed as Geologist with Texaco, Inc., Farmington, New Mexico.

PROCEDURE

Laboratory work involved synthesis of sands of various median sizes and sorting coefficients, impregnation of the sands, measurement of the porosity of each sample, and measurement of grain sphericities in several sands.

Synthesis of Sands

Sands used in the present study were synthesized from sieved fractions of the Lissie Sand (Pleistocene; near Houston, Texas) and the Corrigan Sand (Oligocene; near Corrigan, Texas). Prior to sieving, the sands were washed in dilute hydrochloric acid, and the acid was elutriated out. This procedure removed calcite, most of the iron oxide, and most of the clay. After washing, the sands were dried under heat lamps and then sieved for 15 minutes in a Tyler Ro-Tap shaker.

The various sieve fractions were recombined to form the sands used in the present study. Following the method of Menard (1949), sands were synthesized with lognormal size-frequency distributions and various median sizes and sorting coefficients; a lognormal distribution is uniquely determined by median and sorting coefficient (or standard deviation). Deviations from intended distributions were minimized by choosing median sizes which were equal to geometric midpoints between pairs of sieves previously used to separate the different

size fractions. For each median size, a group of sands was synthesized with simple fractional differences in phi standard deviation (Krumbein, 1936). For lognormal distributions, standard deviations and sorting coefficients (Trask, 1932) are equivalent measures of sorting, and in this paper, sorting coefficient is used.

The only apparent major problem in the synthesis procedure was in connection with sands having appreciable amounts of material of less than 0.044 mm size (325 mesh). This sieve was the smallest size available to the writers, and material passing through it, to be used later in synthesis, was of uncertain size. Microscopic examination of the -325 material, however, showed that it was quartzose silt similar to the material in the overlying sieve, and presumably it was in approximately the desired size range. Not more than 11 percent of -325 material was used in any sample, however, and consequently no sands were synthesized with median sizes of 0.106 and 0.151 mm and sorting coefficients greater than 1.625. The geometric mean size of the -0.044 mm material was taken as 0.040 mm for statistical calculations; 0.040 is geometrically as much below 0.044 as the mean size of the next larger size grade (0.044–0.053 mm) is above.

A total of 250 gm of each sand was synthesized. The material was homogenized in a mechanical mixer which did not break the grains or by stirring with a spoon, which was found to be more effective than a mixer for poorly sorted samples containing large grains.

Impregnation

Porosity measurements were made from sections cut from sand samples impregnated with an epoxy resin. The impregnated sample was prepared by introducing approximately 30 gm of sand plus the resin into a 50-ml centrifuge tube, evacuating, and centrifuging. The sand was placed in the tube by rolling it in paper into a tight cylinder, inserting the cylinder into the tube in a horizontal position, quickly moving the tube to a vertical position, and removing the paper; this procedure prevented size separation of the grains. The resin was then added, and evacuation was carried out in three steps to remove air trapped in the interstices of the sand. Full atmospheric pressure was bled back into the tube after each evacuation. The resin-impregnated sand was then centrifuged at 2500 rpm for 10 minutes. Preliminary tests showed that measured porosities decreased slightly with greater centrifuging time between 2 and 5 minutes, but continued centrifuging between 5 and 20 minutes had no effect on porosity. After centrifuging, the samples were allowed to remain at room temperature for 1 hour to allow the plastic to set and were then cured in a water bath at 70° C for one hour.

Measurement of Porosity

Thin sections were made normal to the core length from the central portions of the impregnated cores. Porosities were then determined microscopically by counting "grain" or "cement" in a 1000-point grid using a point-counting mechanical stage and a microscope equipped with 8× ocular and 25× objective. Preliminary counting of one section showed no appreciable change in measured values of porosity for count lengths greater than 800 points. Some points (no more than 10 in any slide) were directly on the contact between grain and cement, and these were omitted in the porosity calculation. All sections were counted twice, at different times, and the arithmetic average of the two indicated porosities was taken as the true porosity. The average difference between the two porosities measured for each slide was about 1.0 percent, and hence the counting error is believed to be about ± 0.5 percent. The counting error is somewhat greater for samples containing a few relatively large grains, namely those samples of large median size and large sorting coefficient.

Measurement of Sphericity

True sphericity, based on grain surface area (Wadell, 1932), is extremely difficult to measure. Consequently, the writers adopted the simple definition that sphericity of a grain equals the area of the projected image of the grain divided by the area of the smallest circumscribing circle.

Sphericities were measured on two sets of samples: (1) from the thin sections of the best sorted samples for each median size (samples 1, 5, 9, and 14); and (2) from grain

mounts, in immersion oil, of six of the sieve fractions. For each sample, photomicrographs were made, areas of 22–25 grain images were determined by planimeter, and areas of circumscribing circles were calculated from measurements of maximum projected grain diameter. The arithmetic average of the grain sphericities was taken as the average sphericity of each sample.

RESULTS

Relationships between porosity and sorting coefficient are shown in figure 1, which is a composite plot of groups of samples with 4 different median diameters. The general decrease in porosity with increase in sorting coefficient (increase in spread of grain sizes) is well shown. Also, for groups of sands having the same median diameter, the relationship between porosity and sorting coefficient is approximately linear except for very well sorted sands (low sorting coefficient). The highest porosities approach the theoretical maximum of 47.6 percent (Graton and Fraser, 1935).

Curves representing the relationships at different median diameters do not coincide with each other except where the sorting coefficient is very low (sorting is very good). As the sorting becomes poorer, the curves diverge from one another, and it appears that as the median size decreases, the percent porosity increases for the same sorting coefficient. These relationships can be more readily visualized from figure 2, which is a plot of median diameter size against percent porosity at various sorting coefficients. Where the sorting is good, porosity appears to be independent of the actual grain size and is only a function of sorting coefficient. As the sorting becomes poorer, porosity becomes increasingly dependent on median diameter size as well as on sorting.

TABLE 1.—*Relationship between sphericity and grain size*

Median Diameter	Average Sphericity	
	Thin Section	Oil Immersion Mounts
0.335	0.58	
0.213	0.57	0.73
0.151	0.56	0.65
0.106	0.55	0.68
0.075		0.69
0.057		0.65
0.040		0.62

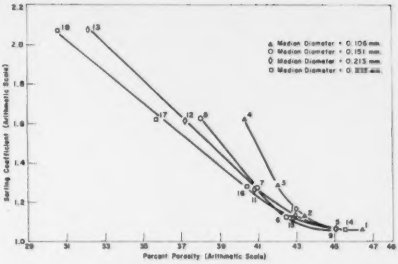
Sorting Coefficient (Arithmetic Scale)
Percent Porosity (Arithmetic Scale)
Median Diameter = 0.106 mm.
Median Diameter = 0.151 mm.
Median Diameter = 0.213 mm.
Median Diameter = 0.335 mm.

FIG. 1.—Relationship between porosity and sorting coefficient of sands with various median sizes.

If all of the sand grains were perfect spheres, porosity should be independent of absolute grain size. Consequently, average sphericities were measured for the best sorted sands of each median diameter and for grain mounts of six of the sieve fractions used to synthesize the sands. Results of the two sets of measurements are shown in table 1. The thin section data presumably represent projected sphericities of randomly oriented grains, for the grains were packed firmly after pouring into the tube, and centrifuging probably did not force them into any appreciable pattern. Data from mounts in immersion oils represent sphericities measured from the maximum projected areas of the grains. These sphericities are higher than those for the randomly oriented grains, which probably indicates that the long and intermediate grain axes are more nearly equal than the short and intermediate axes.

Both sets of measurements show that the sphericity decreases slightly with decreasing grain size. For the thin section data, the variation is extremely small. For the oil immersion mounts the data are somewhat scattered, and consequently the significance of the relationship has been tested statistically. In the oil immersion mounts, the correlation coefficient between sphericity and grain size (using the arithmetic value of median size) is 0.73; using the phi value of median size, the correlation coefficient is 0.71. The critical absolute value of r at the

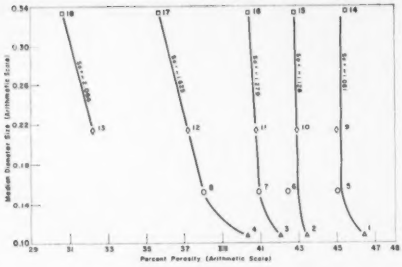

FIG. 2.—Relationship between porosity and median size of sands with various sorting coefficients.

95 percent confidence level is 0.73. The regression coefficients are 0.43 using arithmetic median sizes and -0.029 using phi values. The conclusion is that sphericity decreases with grain size and that the scatter shown by the various grain mounts is sampling variability. For this reason, the increase in porosity with decreasing grain size is attributed to decrease in sphericity and consequent poorer packing.

SUMMARY

The following general conclusions can be drawn from the present study.

1.—Well sorted sands are more porous than poorly sorted ones. The relationship between porosity and sorting coefficient is approximately linear except where the sorting is very good, in which case percent porosity increases more markedly than would be predicted from a linear relationship.

2.—In well sorted sands, porosity is independent of actual grain size.

3.—In poorly sorted sands, the finer ones are more porous than the coarser ones, and the effect becomes increasingly apparent as the sorting becomes poorer. This relationship is presumably caused by lower sphericity of the smaller grains.

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PLATE 1, FIGS. 1–6, SEPTEMBER, 1961

NOTES

A NON-LAMINATED MINIATURE SAMPLE SPLITTER[1]

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ABSTRACT

The chutes of this sample splitter are milled from a solid brass bar, and the receiving pans are made from plastic blocks. A revolving tray is fitted above the hopper to eliminate the personal element in feeding the sample into the splitter. Rigidity and freedom from contamination results from this form of construction.

INTRODUCTION

Since Otto (1933) first described his miniature sample splitter, various alternative methods of construction have been suggested. For example, Skolnick (1959) has advocated the use of glass plates cut diagonally and clamped in a simple press, while McKinney and Silver (1956) have described a joint-free sample splitter folded from a single sheet of aluminium. The laminated form of splitter suffers from the possibility of material lodging in crevices resulting in loss or in contamination of subsequent samples. The folded form avoids this problem but would appear to be somewhat flimsy, especially in the smaller sizes.

These objections have been overcome by milling the chutes in a solid brass bar. Rigidity without the necessity for clamping is thus achieved and there are no crevices where particles can become trapped. The complete splitter and the component parts are shown in figure 1.

DETAILS OF CONSTRUCTION

The milling of the brass bar was carried out on a 3½-in centre lathe using a 2½-in diameter plain milling cutter 0.051 in (1.3 mm) thick mounted on a 1-in diameter arbor carried between the chuck and the tail-stock. The bar was clamped in a vise bolted to the cross slide and packed up until the top of the bar was 7/16 in below the centre line of the lathe. In this way the slope of the chute was kept at a steep angle. Eight slots, 5/16 in deep, were cut at intervals of 0.132 in using the lead screw index. Eight similar slots were made on the opposite side of the bar by traversing the cross slide to the back of the lathe, again using the lead screw index to position the cutter.

In the absence of a lead screw index it is possible to position the cutter using feeler gauges of the appropriate thickness and a small hand lens. This can be done by pressing the gauge against the cutter and adjusting the position of the block until the edge of the preceding slot is coincident with the edge of the gauge. Using this method, partitions as thin as 0.005 in have been cut, but because of the difficulty of setting the cutter with the required degree of accuracy, it is preferable to adopt the greater wall thickness of 0.015 in as used in the present splitter. Subsequent measurement of the wall thickness showed a mean value of about 0.012 in, probably caused by a slight wobble in the cutter. The use of a milling machine or a precision lathe might eliminate this.

The receiving pans were milled from ⅜-in thick plastic (Perspex, Lucite) blocks using a ⅜-in diameter end mill. The surface of the block was inclined at approximately 15° to the length of the cutter.

The revolving tray was made from a length of brass tube plugged at each end with short lengths of brass rod. Part of the wall was then cut away to form the tray.

The dimensions of the various parts are shown in figure 1. It should be noted that none are critical but that the internal length of the tray, the space between the tray supports, and the distance between the

[1] Manuscript received June 6, 1960.

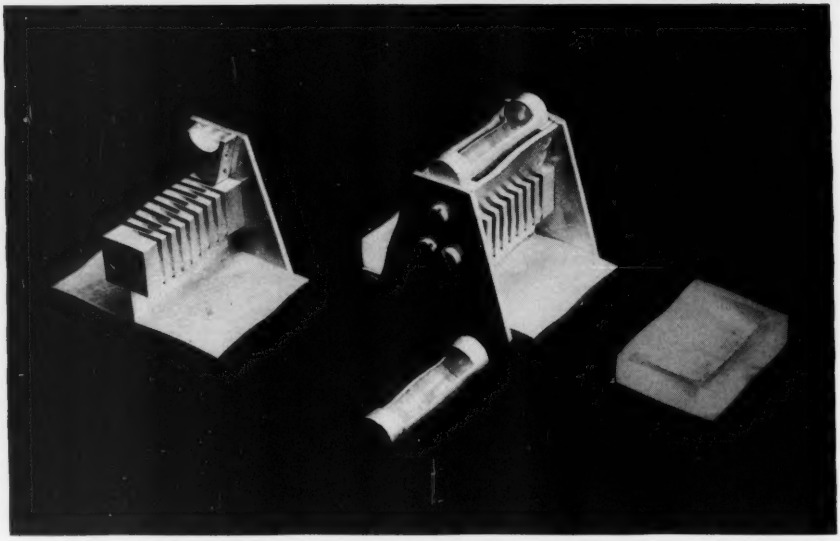
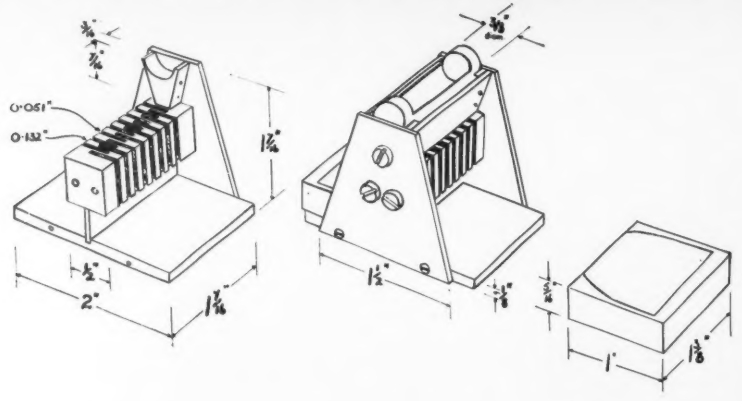


FIG. 1.—(above) The complete splitter together with the revolving tray, a receiving pan, and a partially dismantled splitter showing the rigid splitter bar.
(below) Line diagram to show the dimensions of the complete splitter and component parts.



outer edges of the end chutes should be the same. This prevents loss of grains in the supports. Furthermore, the tray should rest on its supports with a minimum of end play.

The cost of the materials is negligible, while the time taken to construct the prototype was about two man-days.

METHOD OF USE

The maximum capacity of the tray is approximately 1 gm of quartz sand. This is adequate for most heavy mineral concentrates, but samples required for shape analysis, for example, should first be reduced to approximately 1 gm by hand quartering or preferably by a rotary splitter. The sample is placed in the tray which is then rotated by a rolling motion of the fingers. A good polish on the tray and its supports will facilitate the feeding of the sample into the

splitter. Lubricants should not, of course, be used. Small samples should be evenly dispersed along the tray. A slight tap will dislodge any particles perched on the partitions.

ACCURACY OF SPLITTING

Since the device described here is exactly similar in principle to Otto's "Microsplit" it is unnecessary to reproduce the numerical results of tests on the splitter. The slight variation in wall thickness mentioned above has no observable effect on the accuracy of the splitter.

CONCLUSIONS

The chief advantages of this type of construction are:

1.—the rigidity of the splitter bar,
2.—the absence of thin unsupported partitions liable to damage.
3.—the absence of crevices where mineral grains might lodge,
4.—the comparative simplicity of construction,
5.—the elimination of the personal error in feeding the sample into the chutes, provided the sample is spread along the length of the revolving tray.

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EARLY DIAGENETIC DOLOMITE ASSOCIATED WITH SALT EXTRACTION PROCESS, INAGUA, BAHAMAS[1]

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ABSTRACT

On the island of Inagua in the West Indies large quantities of salt (NaCl) are produced by the evaporation of sea water. Samples of the precipitates taken at various stages in the evaporation process do not contain the mineral dolomite in discernable quantities. Clear, pale-pink dolomite rhombohedra up to 0.14 mm long are present in the sludge dump waste dredged from the evaporation pans less than 8 years ago. The rhombohedra are dispersed between crystals of other substances but give no clue as to their specific origin.

INTRODUCTION

On Great Inagua, southernmost of the Bahama Islands in the West Indies, salt is extracted from sea water commercially by natural solar evaporation. West India Chemicals, Limited, a subsidiary of the Morton Salt Company, conducts the operation and exports in excess of 175,000 tons of salt (99% NaCl) a year. In this process normal sea water at a temperature of 70–75° (average) is routed through a series of shallow reservoirs 14 to 30 inches deep to permit the settling of solids. The arid climate causes the sea water to evaporate rapidly, and as the specific gravity increases the brine is transferred closer to a series of final evaporative pans. Salt is precipitated from approximately 12 in. of water in the pans when the specific gravity reaches 1.220 at a temperature of 90 to 105° F; the final brine is drained away at 1.245 to prevent magnesium contamination. One hundred gallons of sea water reduces by evaporation to 10 gallons of brine with a specific gravity of 1.210 which yields approximately 23 lbs of NaCl.

The process described above and illustrated in plan view in figure 1 was utilized as a large natural outdoor laboratory to study

[1] Manuscript received April 28, 1960.

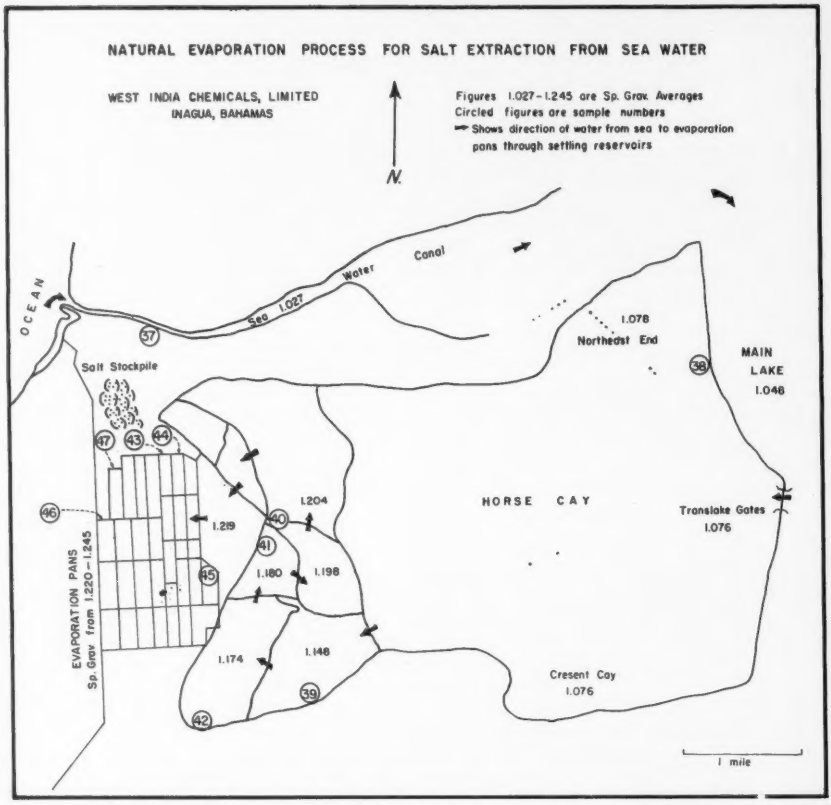

FIG. 1.—Plan view of West India Chemicals, Limited salt extraction process showing the water route through the settling reservoirs and evaporation pans.

carbonate precipitation in conjunction with evaporite conditions. By analyzing the mineral precipitates at various stages during evaporation it is possible to study in detail the transformation from normal sea water to solids as they occur naturally. The writer used these facilities in an attempt to study the direct precipitation of the mineral dolomite.

SAMPLING AND ANALYSIS

In the latter part of 1957 glass sample bottles with plastic caps and plastic lined sample sacks were used by Robert Fanfant Chief Chemist, West India Chemicals, Limited, to procure air dried samples of the bottom materials at the numbered locations in the settling reservoirs and evaporation pans shown on the map in figure 1. An additional sample was taken from the base of the sludge (waste) dump, material dredged from the pans about 8 years ago.

Mr. Fanfant, with the permission of A. W. Erickson, Jr., President, West India Chemicals, Limited, also provided chemical analyses of the brines at each sampled location as well as additional information on the sea water, bottom mud, and waste sludge to further this investigation.

By using the samples in conjunction with

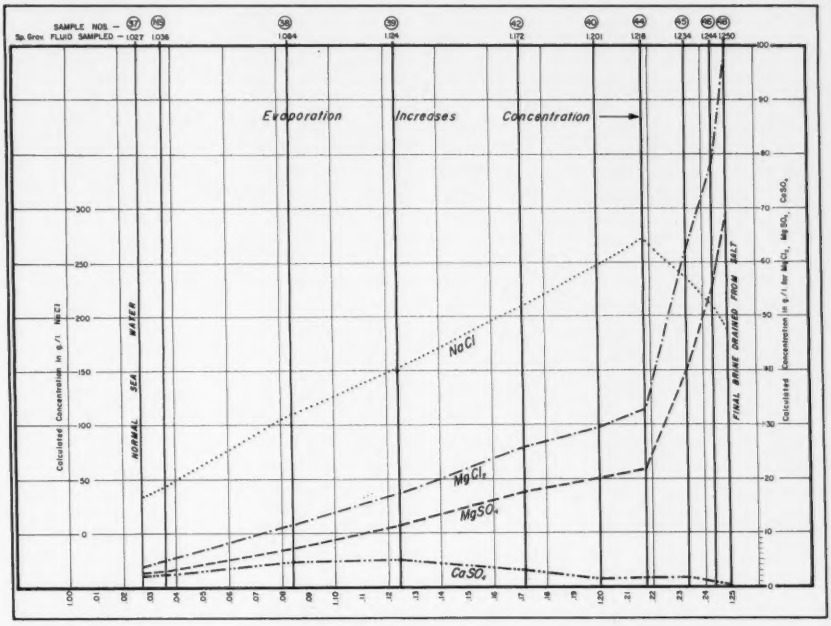

FIG. 2.—Profile of sea water chemistry showing the changes in concentration as the evaporation reduces it to brine.

the chemical analysis of the brine from which they were derived, it is possible to plot (fig. 1) a chemical profile of the precipitation from normal sea water to final brine. Binocular and petrographic descriptions supplemented by X-ray analysis were used to describe and identify the minerals in the samples.

Chemical analyses were obtained on 10 samples of Ca^{++}, Mg^{++}, SO_4^{--}, and Cl^- concentrations in the brine and recorded in terms of grams per liter. Theoretical combinations of these salts were recorded (table 1) in the same manner and plotted as a normal sea water to final brine profile (fig. 2).

CONCLUSIONS

The analyses failed to reveal dolomite in perceptible quantities in any of the samples taken from the mineral precipitates in the

TABLE 1.—*Water chemistry*

	Sample No.	Normal Sea Water 37	NS	38	39	42	40	44	45	46	Final Brine 48
	Specific Grav.	1.027	1.036	1.084	1.124	1.172	1.201	1.218	1.234	1.244	1.250
Analyzed concentration g./l.	Ca^{++}	0.595	0.629	1.320	1.450	1.240	0.49	0.540	0.490	0.02	tr.
	Mg^{++}	1.400	1.866	4.286	6.653	6.820	11.50	13.10	24.14	30.9	39.5
	SO_4^{--}	3.123	3.880	8.958	12.39	12.33	17.01	18.42	34.1	44.0	56.1
	Cl^-	21.65	27.65	64.00	97.60	98.80	170.8	189.7	189.3	186.4	186.3
Theoretical combination g./l.	$CaSO_4$	2.01	2.13	4.47	4.91	3.21	1.63	1.82	1.84	0.08	—
	$MgSO_4$	2.14	2.97	7.00	11.19	17.32	19.98	21.83	40.0	55.20	70.35
	$MgCl_2$	3.79	4.95	11.24	17.19	25.57	29.10	32.73	61.5	77.8	99.4
	NaCl	34.09	43.27	108.5	152.9	209.1	246.5	272.8	235.2	211.9	184.8

settling reservoirs or in the evaporation pans, suggesting that increased salinity and concentration alone were not sufficient to cause the precipitation of dolomite in brine with a specific gravity less than 1.245.

Clear, pale-pink, dolomite rhombohedra up to 0.14 mm long were found disseminated with subhedral aragonite and calcite in the sludge dump sample which has been above the water level approximately 8 years. The rhombohedra are intricately associated with the other evaporite minerals but give no clue as to their specific origin. No two dolomite crystals were observed in close proximity to each other; they appear to be dispersed and may have formed in interstitial voids between the existing crystals. Although the dolomite is most certainly younger than the enclosing matrix there is no direct evidence to substantiate that it formed as a replacement product after aragonite, vaterite, or calcite. In all probability the rhombs of dolomite formed in minute voids where magnesium rich (see table 1) brine accumulated or filtered through the waste sludge as new material was dredged onto the top of the dump. Future investigations of the dump samples on a chronological basis, literally month by month or year by year, will probably show a successive increase in dolomite with time, both as a replacement product and as a primary precipitate.

ACKNOWLEDGMENTS

The author wishes to acknowledge the assistance of Dr. T. R. Walker, University of Colorado, for providing X-ray analyses and Dr. R. L. Folk, the University of Texas, for his comments and criticism of the manuscript. I also wish to express my sincerest appreciation to Robert Fanfant for obtaining the samples as well as providing all of the chemical analyses and to the West India Chemicals, Limited for permission to publish this material.

THE FLOTATION PHENOMENON—A BY-PASSING AND SORTING PROCESS[1]

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ABSTRACT

The tidal rivers of Cape York, Australia, transport well-sorted, angular quartz sand into the Gulf of Carpentaria by flotation. The combination of the environmental conditions causing this process are firstly, the peneplanation of the coastal area; secondly, the tidal rivers which run upstream for several miles; thirdly, the semi-tropical climate; fourthly, the lateritic and bauxitic soil containing a high percentage of very angular, quartz grains. The flotation phenomenon acts as a by-passing and sorting agent, and large amounts of clean sand are transported in this manner in various parts of the world, where other environmental combinations may permit the flotation of sand.

During the dry season patches of "floating sand" were found to drift in the Wenlock River, Queensland, to the Gulf of Carpentaria. The size of the quartz sand patches varied from a few millimeters up to approximately 10 cm in diameter, and their thickness was that of one grain. The sand rafts consisted of a loose network of very angular quartz constituents with the smallest possible space between the individual particles. Sorting was generally good; the average grain size was between 2 and 3 mm in diameter. No other material (for example organic matter or a film of oil) that could explain the floating of the sand was present. The weight of the sand caused a narrow raised rim of the river water to surround the patches. The sand rafts passed the observer quite irregularly. Up to six patches drifted by in one minute. The floating sand was

[1] Manuscript received June 21, 1960.

considerably unstable and when a patch was touched in the center all the grains sank.

To find an explanation for the flotation process the environmental conditions of the western part of Cape York have to be considered.

During the dry season (April-December) many of the numerous westward running stream beds contain only little freshwater near the Gulf. Further upstream they may be completely dry. The western part of Cape York peninsula is semi-peneplaned, and large areas are just a few feet above sea-level. The rivers, therefore, have very low gradients and the tidal water moves upstream for a few miles. The heavier salt water moves as a wedge underneath the freshwater, while the latter continues to flow as a relatively thin layer down-stream.

The top 30 to 50 ft of the soil is lateritic to bauxitic which has been formed by chemical weathering of sedimentary rocks, for example sub-graywackes. Some of the residual soil contains up to 60 percent very angular quartz grains. Along the river beds and especially on flood plains, the quartz accumulations formed large-sized deposits, such as levees.

It is unknown to the writer whether the flotation phenomenon is possible during the monsoonal rains (this area has more than 60 inches of rainfall within a few months time). The levees certainly were formed during the wet seasons, and it seems possible that at times they may supply quartz sand for the flotation process. Especially along the flood plains the water will move slowly as a sheet without much turbulance, and sand rafts could be moved by flotation to the deeper water of the Gulf.

The rising tidal water is able to pick up the sun-dried quartz just as it would lift up a leaf, for example. The sand remains afloat as long as the grains are dry. McKelvey (1941) gave a summary on the experiments performed to explain flotation. It has been found that the surface tension of water can support quartz grains up to 5 mm in diameter.

The type of source area is of importance. If the original subgraywacke contained 20 to 30 percent quartz, the chemical weathering will remove all the relatively soluble constituents and concentrate the iron and aluminum oxides, quartz as well as some other less important material.

The residual soil, therefore, may contain twice or three times the amount of quartz as compared to the original rock before chemical weathering. The transportation path of the individual grains during the chemical weathering process has a vertical and a possible horizontal component. Although a long period is required, this process does sort and concentrate the material on hand without the grains undergoing mechanical wear. It seems that the formation of tropical soils by chemical weathering of a basalt, for instance, would not be accompanied by a concentration of quartz or any other mineral that could serve as a source for flotation.

As is well known, a semi-tropical to tropical climate and a very low topography to prevent rapid erosion are required for the formation of a residual soil.

The advance of peneplanation is accompanied by a decrease of river gradients and therefore an increase of the available source area for the land-inward flowing tidal waters.

The material transported by flotation can by-pass the sediments moved by the more commonly known processes—that is, traction, saltation, and suspension.

Behre (1926) estimated that in one hour 130 tons of material were transported by flotation by the Ohio River. Ladd (1898) calculated that the Merrimac River transported 1000 tons of sand a year by the same process.

Based on such available information, the writer (Wolf, 1958) had suggested that a well-sorted, angular quartz-sand deposit may be the result. But Emery (1960, personal communication) writes "that sand flotation is common but always only a minor form of transportation. . . . I know of no deposits formed uniquely of floated sands, and doubt if any exists. Instead, the floated sands are deposited with others that are so abundant as to mask the floated sands."

Many of the tidal rivers flow into the ocean waters without creating turbulance which is confined mainly to the beach. There is, therefore, in many cases no major barrier which would prevent the transportation of material by flotation to reach the

deeper parts during calm periods. At other times the rough sea will cause the sinking of the sand near the shore, as described by Emery (1945).

It may be concluded that the following conditions gave rise to the flotation process at Cape York, Queensland: firstly, peneplanation of the coastal area; secondly, the inland-moving tidal streams; thirdly, the semi-tropical, monsoonal climate; and fourthly, the lateritic and bauxitic residual soil containing up to 60 percent very angular quartz grains.

Other environmental combinations may be conducive to flotation. For example, McKelvey (1941) mentions that off-shore winds can blow sand onto the water surface of a sea and move the sand seaward. Slowly rising tidal waters and peneplanation, for instance, are therefore not always necessary. Emery mentions that he has observed in many parts of the world the flotation phenomenon and states that "these areas include highlands—not all are lowland peneplained areas."

Also, great angularity of the sand grains is not a necessity although angular particles float easier (McKelvey, 1941). It is unknown to the writer whether this phenomenon could result in the separation and consequent sorting of sand grains of different angularity by the flotation process, Emery writes that he has "been able to detect no greater angularity of floated grains than of source sand, except for the preference for shell fragments and biotite."

Pettijohn (1958, personal communication) states that the flotation process has been observed by others many times. Emery also has seen the flotation of sand in Japan, Guam, Hawaii, Bikini, New Zealand, Israel, Egypt, Saudi Arabia, Bermuda, France, and and many other places. He mentions that "flotation of sand is quite common but is it merely overlooked by all but a few sedimentologists."

Acknowledgments.—Grateful acknowledgments are made to Dr. K. O. Emery and Dr. F. J. Pettijohn for their information.

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A NOTE ON THE REPLACEMENT IN JOINTS[1]

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ABSTRACT

Spectrographic analyses were run on a solution enlarged joint in the Salem Limestone of Indiana. The replacement of calcium carbonate is most active at the wall of the joint and decreases away from it. It is suggested that the process of replacement in limestone may be facilitated by the joint pattern, and that the joints may offer an easy path for the solution, which simultaneously dissolves the carbonates and replaces them.

DESCRIPTION OF SALEM LIMESTONE

The limestone is massive, thick bedded, with well-developed cross bedding, joints, and stylolites. The color is light or dark gray; and in a few places buff or drab. In thin lenses the color may be different.

The top of the formation grades into somewhat poorly dolomitic and argillaceous St. Louis Limestone, which is calcium carbonate of a considerably lesser purity.

[1] Manuscript received July 15, 1960.

TABLE 1.—*Spectrographic analyses of Salem Limestone joint*

No.	Distance in cm. from joint	MgO%	Fe%	Al_2O_3%	SiO_2%	SrO%	TiO_2%
1	0	8.3	0.60	2.30	7.50	0.020	0.17
2	6.5	7.4	0.30	1.10	4.0	0.018	0.070
3	11.4	7.3	0.35	1.20	5.3	0.014	0.080
4	13	6.8	0.35	1.10	4.7	0.016	0.075
5	14.5	7.0	0.30	1.10	4.0	0.013	0.085
6	17.8	6.5	0.30	1.10	4.2	0.017	0.080
7	21.5	6.4	0.28	1.10	4.0	0.016	0.075
8	25.3	5.7	0.27	1.0	3.7	0.018	0.07
9	30.3	4.5	0.20	0.50	3.50	0.018	0.10

The Salem Limestone is granular, and pseudo-oolitic, the grains consisting mainly of fragments of fossils with very abundant foraminifers and ostracods. At the base of the formation abundant fenestellid bryozoans are found. At the top and along the joints silicified specimens are found. Crinoid fragments, gastropods, brachiopods, *Pentremites*, and pelecypods are the most common types. Fossils are more abundant at some localities than at others. In some places only micro-fossils are found; in others only large bellerophonoid gastropods are found.

Most of the rock is of granular texture. The grain size varies from fine to coarse. There is evidence of sorting. No dolomite was found, but silicification along the joint pattern is common.

The rock has been called spergenite, and its chemical analyses have shown it to be relatively pure carbonate with calcium carbonate as high as 98 percent. (71.6 to 98.27 percent; Perry and others, 1957; Loughlin, 1929.)

The joints run in an east-west direction. They are nearly vertical tension joints. Most of them are solution enlarged at the top of the formation. The samples were collected at such an enlargement in Hoadley Quarry, east of Arlington, 2 miles northwest of Bloomington, Indiana. Many of the surfaces of the joints are well silicified and weathered.

CHEMICAL ANALYSES

The spectrographic analyses represent 9 samples taken away from the joint surface for a distance of 30.3 cm. The analyses are given in table 1.

Since many of the joints in the formation cut across a number of large stylolites, they were formed in the post stylolitic time. Four small stylolites were found running across the area studied and were found not to effect the chemical analyses.

The percentage of strontium changes very little from one sample to another (fig. 2). Titanium and aluminium show a very marked decrease (titanium about 60 percent, aluminium about 50 percent) from the joint surface up to the first 6 cm (figs. 1 and 2). From there on titanium concentration practically does not change while aluminium shows a small but steady decrease.

Iron (fig. 1) shows a slow but considerable

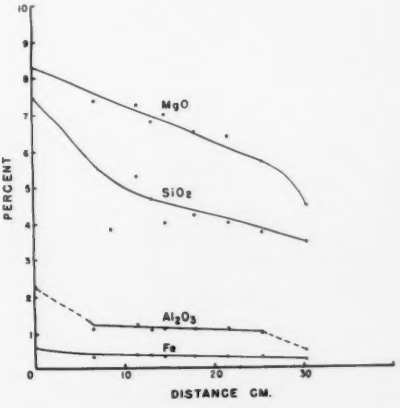

FIG. 1.—The graphic representation of the percentage of MgO, SiO_2, Al_2O_3, and Fe plotted against the distance in cm.

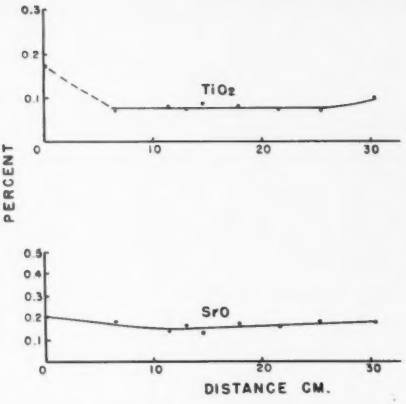

FIG. 2.—The graphic representation of the percentage of TiO_2 and SrO plotted against the distance in cm.

linear decrease away from the joint (about 50 percent).

The greatest changes have taken place with magnesium (fig. 1) and silica (fig. 1). Magnesium concentration unlike silica shows a steady, linear decrease. Although silica's very marked decrease is unmistakable, there are several points which do not fall on any curve one wishes to draw in this case. One can almost conclude that from somewhere about 6 cm away from the joint, silica concentration is almost unchanged. From this analysis one cannot conclude how the silicon changes more than 6.5 cm away from the joint surface.

DISCUSSION

It appears that the replacement was caused by the ground water, which caused joint enlargement and composition changes around the joint; both phenomena must have occurred simultaneously. Waters along joints are usually not stationary, but are moving rather rapidly relative to the pore fluids. The water may have been derived from the rock itself and, since the porosity of limestone is high, the amount of water must also be high. Loughlin (1929) found that as much as 1.19 percent of all the constituents present was water.

The replacement shown here may offer a possible clue to the direction of the replacement in the rock and also the possible role of the joints in the process. It is not silicification alone that took place but Mg, Fe, and Al seem to have "migrated" into the rock. The degree of alteration is dependent upon the distance from the joint surface.

Three processes come to mind which may explain the composition changes:

1.—The leaching out of $CaCO_3$, leaving the more insoluble impurities behind.

If the water in the joint is in a moving state, then the different ion concentration can be maintained, and diffusion caused by the concentration gradient can exist. It is expected that in the Salem Limestone, owing to its high porosity and large joint system, the water must have had rapid flow. The disequilibrium might be high enough to allow the lime to enter from the saturated pore solution into the joint system fluids to be moved away. This may explain the solution of lime.

This process would have left behind a marked change in the rock structure. The pore spaces would have been enlarged, and this is not the case as seen in figure 3. Also the remarkably high concentration of magnesium is thus difficult to explain.

2.—Another process is a mechanism by which the impurities are brought up by the migration of the joint fluids. Here it is assumed that the replacing materials have their source elsewhere and that they were brought in and concentrated in joint fluids, and then migrated into the rock. The replacement in this process might be controlled by either concentration gradients or the actual rates of migration of ions.

This process has the difficulties of explaining how the impurities are carried in solution if they are insolubles, and in what form? Magnesium concentration is also unexplainable here.

3.—The third possibility is that the process is one of diffusion and not of flow. The surface of the rock at the joint contact can be considered to be a different rock. All elements analyzed for are at a higher concentration at this contact zone. One can almost call these different phases.

The order of decreasing mobility of oxide "molecules" is given by Turner (1948, p. 52) as: H_2O, CO_2, CaO, MgO, FeO, Fe_2O_3, SiO_2, Al_2O_3 (those only relevent to the lime-

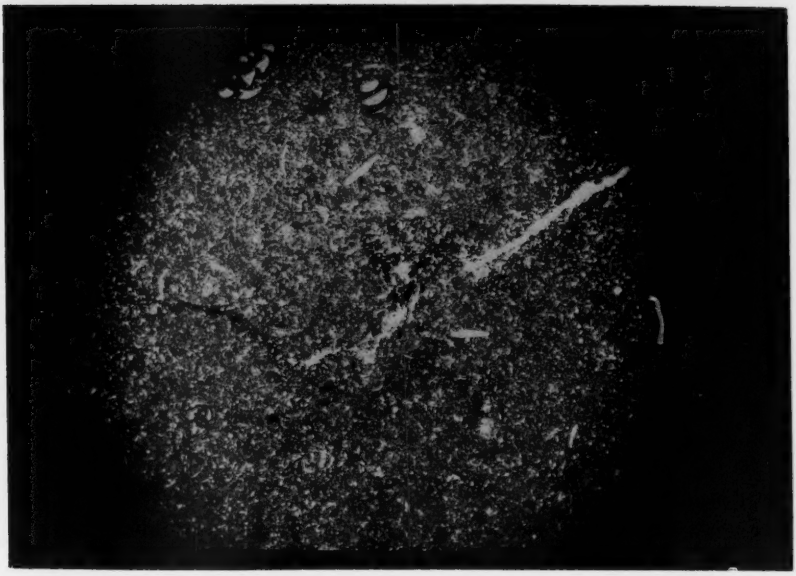

FIG. 3.—Photomicrograph of a thin-section cut parallel to the surface of the joint. The black lines are the small stylolites with the "typical" insolubles. The white elongated crystal is the small calcite vein. (×15).

stone are quoted). In ionic form the following is the order of decreasing mobility: CO_3^-, Ca^{++}, Mg^{++}, Fe^{++}, Al^{+++}. Probably, as Turner points out, important in the velocity of diffusion is a concentration gradient set up between moving and stationary solution. Thus, this offers even more probability for diffusion of lime from the pore solution into less saturated water in the moving system.

CO_2, the main component of the limestone, can enter the pore solutions; probably lower pH conditions existed, which in contact with the alkaline walls of the joint caused precipitation of Fe.

CONCLUSION

In conclusion, one may say that none of these processes really adequately explains the phenomenon; very possibly all mechanisms were in operation at the same time.

One can say with assurance, however, that the joint controls the increase of the concentration of elements relative to the concentration of $CaCO_3$ around it.

The causes of these distributions lie partly in the joint complex and partly in the composition of the rock which, of course, is itself to a considerable extent a product of the chemical change.

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AUTHIGENIC TOURMALINE AND ZIRCON IN THE VINDHYAN FORMATIONS OF SONE VALLEY, MIRZAPUR DISTRICT, UTTAR PRADESH, INDIA[1]

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In the course of a detailed sedimentological study of the rocks of the Vindhyan formations of the Sone Valley, Mirzapur District, Uttar Pradesh, India, authigenic overgrowths have been observed on the tourmalines and zircons in the arenites from these formations (pl. 1).

The Vindhyan formations (late Pre-Cambrian to early Palaeozoic?) of the Sone Valley are made up of the Lower Vindhyans (Semri Series) and the Kaimur Series, that is the lowermost series of the Upper Vindhyans. The rocks of the Semri Series are mainly calcareous with bands of sandstones and shales. The Kaimur Series is chiefly arenaceous and is made up of sandstones and quartzites with subordinate shales.

Authigenic growths on tourmaline have been reported by Alty (1933), Krynine (1946), Gundu Rao (1952), and Gokhale and Bagchi (1959) among others. The present author reports such occurrences in the arenites of the Vindhyan formations for the first time.

Authigenic tourmaline is present in all the arenaceous sediments of both the Semri Series and the Kaimur Series, while the authigenic growth in zircon has been observed only in the Upper Quartzites of the Kaimur Series. Of the total heavy residues the percentages of tourmaline and zircon vary between 26.55 and 65.55 percent and 8.58 and 30.05 percent respectively in the various horizons.

The tourmalines have been classified on the basis of color, shape, and inclusions into various types. The brown and black varieties are the predominant types. Authigenic growths are present on almost all the types of tourmaline grains but are most common on the brown variety. The percentages of the various types of tourmaline are shown in table 1.

The overgrowths are colourless to light blue and are generally small, ranging from 0.02 to 0.12 mm in length. According to Alty's pyroelectric tests, these overgrowths are restricted to the antilogous pole, which is characterised crystallographically by the faces r ($10\bar{1}1$) and m ($10\bar{1}0$). Their R.I. is lower than that of the nucleus, varying from 1.61 to 1.63. The same characters have been observed in my material.

According to Krynine (1946), sedimentary authigenic tourmaline generally forms in cold water at the bottom of the sea, penecontemporaneously with the including sediment. The tourmalines reported in this paper occur firstly, in shallow water marine sediments: glauconitic sandstone and secondly, in the fluviatile deposits (Kaimur Series), formed along flood plains in inter-

[1] Manuscript received September 7, 1960.

EXPLANATION OF PLATE 1

Authigenic growth in Tourmalines:

FIG. 1, 4, 8, ×870 in Upper Quartzites
FIG. 2, 3, ×870 in Green Scarp Sandstone
FIG. 5, 13, 15, ×180 in Sandstone Dyke
FIG. 6, 14, ×180 in Dhandraul Quartzite
FIG. 7, 9, ×180 in Lower Quartzite
FIG. 10, 11, ×180 in Red Scarp Sandstone
FIG. 12, ×180 in Glauconitic Sandstone

Growth phenomena in Zircons of Upper Quartzite: Figs. 16 to 20 ×870.

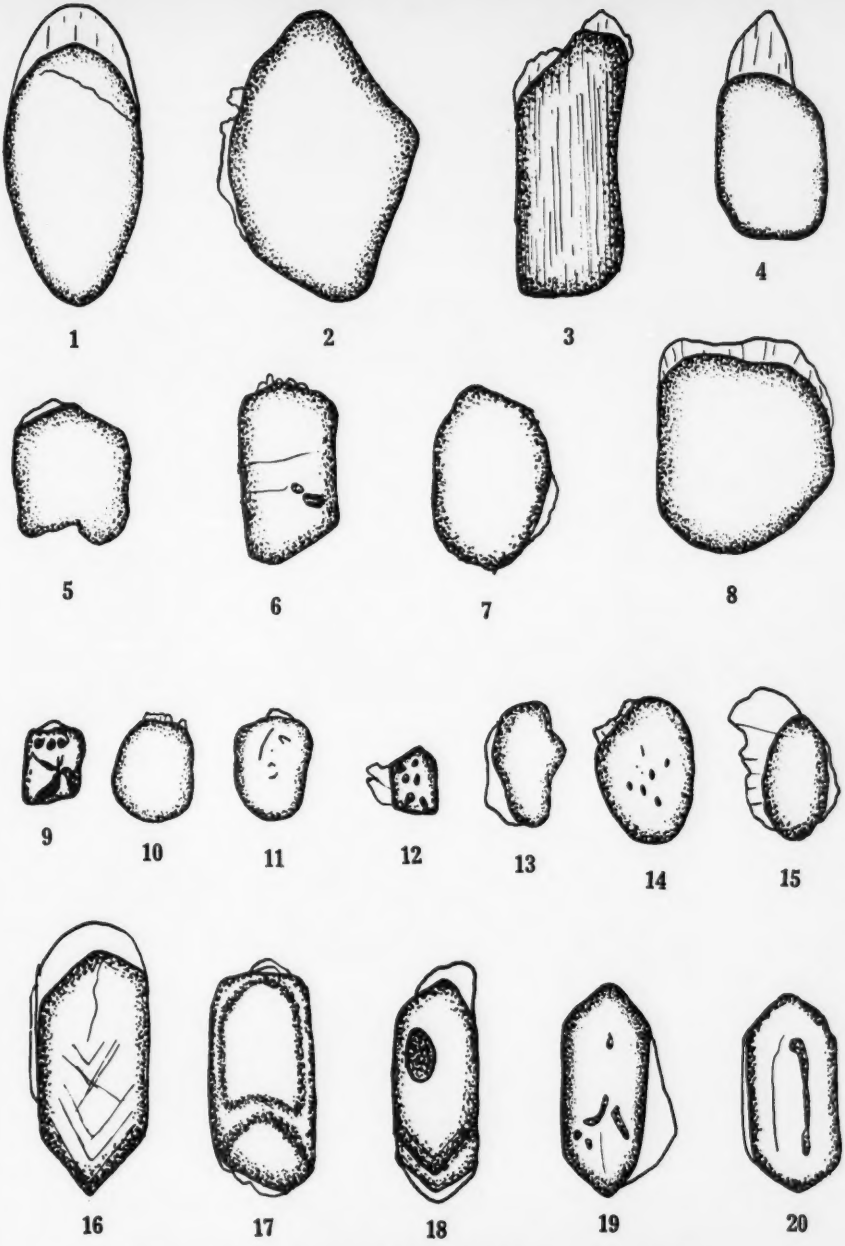

PLATE 1.—Camera lucida sketches of anthigenic tourmaline and zircon.
See facing page for explanation.

TABLE 1.

	Formations	Varieties of Tourmaline	Percentage of Each Variety	Percentage of Authigenic Tourmaline in Each Variety
Kaimur Series	Dhandraul Quartzite	Black	40.3	66.7
		Brown	38.6	41.7
		Blue	6.2	—
		Green	14.9	33.3
	Green Scarp Sandstone	Black	20.9	25.00
		Brown	54.2	45.9
		Blue	8.8	33.3
		Green	14.2	27.5
		Colourless	1.9	—
	Red Scarp Sandstone	Black	19.6	33.3
		Brown	64.4	67.1
		Blue	—	—
		Green	16.0	33.3
	Upper Quartzite	Black	17.5	—
		Brown	56.1	16.77
		Blue	10.2	11.11
		Green	16.2	8.4
	Lower Quartzite	Black	25.2	33.3
		Brown	58.7	16.7
		Blue	—	—
		Green	16.1	24.2
Semri Series	Sandstone dyke	Black	24.1	6.4
		Brown	58.6	11.11
		Blue	5.1	7.7
		Green	17.2	6.9
	Glauconitic Sandstone	Black	28.1	50.00
		Brown	58.4	30.9
		Blue	4.6	12.5
		Green	6.2	25.0
		Colourless	0.8	—

montane basins subject to seasonal rainfall and semi-arid conditions. The environments of the Vindhyan Sediments, however, are more akin to the observations put forward by Takahashi (1955), who states, "that glauconitisation takes place at not lower than 15° C and a salinity of not less than normal sea water."

The growth phenomena in zircon is described from coarse grained arenites by Butterfield (1936) and Smithson (1937) among others. Poldervaart and Ecklemann (1955), however, have reported the growth phenomenon in the zircons of the Authochthonous granites resulting from the transformation of the sediments of Beartooth Mountains of Montana-Wyoming. The zircons exhibit growth phenomena which bridge the gap between the predominantly rounded zircons of siliceous sedinents and the predominantly euhedral zircons of intrusive granites.

ACKNOWLEDGMENT

I am grateful to Dr. R. C. Misra, Professor and Head of the Department of Geology, Lucknow University, for guidance and facilities for work. The present study forms a part of a comprehensive project, "The Sedimentological Study of the Vindhyan Formations."

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 31, NO. 3, PP. 485–486
SEPTEMBER, 1961

REVIEW

A Traverse of Post-Avis Cisco Rocks, Brazos Valley, North Central Texas, Society of Economic Paleontologists & Mineralogists, Permian Basin Section, April, 1960. General Chairman, Thomas Hambleton, Field Trip Leader, L. F. Brown, Jr. with Supplementary papers by D. A. Myers, H. L. Williams, and R. W. Rall. Also enclosed is a Bureau of Economic Geology Report of Investigation No. 41 entitled, "Stratigraphy of the Blach Ranch—Crystal Falls Section (Upper Pennsylvanian) Northern Stephens County Texas" by L. F. Brown, Jr. and a Bureau of Economics Geology's, "Preliminary map of the Blach Ranch and Breckenridge limestone of Young and Jack Counties, Texas" by L. F. Brown, Jr. 66 pages, $8\frac{1}{2} \times 11$ inches, 25 figures, two plates, Copies may be ordered from Microviolite Co., 202 North Big Spring, Midland, Texas. Price $6.50.

It had once been said by a geologist who worked this outcrop belt that "North Central Texas Pennsylvanian geology is deceptively simple especially at 50 miles per hour." A corollary might be "the complexity of the geology varies inversely with the velocity of the observer." However, on this trip the speed was reduced in order to closely observe the complexity of the stratigraphy in an area that has been questionably mapped until the Bureau of Economic Geology completed their detailed study.

This publication is far more than a guidebook of a very enlightening complex area but is in fact a well edited assemblage of five papers covering north central Texas and a correlation with the Eastern Shelf and Central Basin Platform of west Texas.

The papers are grouped into two general categories:

Road Log:

L. F. Brown, Jr., "Introduction to Load Logs" with correlation chart and a chart of inferred depositional conditions." 5 pages.

Road Log, first day—from Cisco, Texas. Saddle Creek Limestone to part of the Harpersville Formation near Breckenridge, Texas. 18 pages.

Road Log, second day—starting at the Breckenridge Limestone contact and then observing the Blach Ranch Limestone formation. Studying exposures of the Thrifty Formation and observing an excellent post Blach Ranch Sandstone channels. 23 pages.

General:

Donald A. Myers, "Stratigraphy of the Cisco Group, Wayland Quadrangle, Stephens and Eastland Counties, Texas 13 pages.

Harold L. Williams, "Cisco Series on the Central Basin Platform." 4 pages.

Raymond W. Rall, "Cisco Series of the Eastern Shelf and Midland Basin." 3 pages.

L. F. Brown, Jr., "Stratigraphy of the Blach Ranch—Crystal Falls Section (Upper Pennsylvanian), Northern Stephens County, Texas." Bureau of Economic Geology, University of Texas Report of Investigations. No. 41, 1960.

Reliable information about north central Texas on paleontology, stratigraphic relationships and sedimentary processes has not been available for much of the Pennsylvanian outcrop belt. Most of the surface studies were of a reconnaissance nature and interest in surface geology decreased after oil geologists mapped the surface structures years ago. A more thorough understanding of the outcropping rocks is most desirable for stratigraphic framework, detailed zonation of invertebrates, spores, pollen, and investigating environmental control of these flauna and flora which will benefit subsurface geology. This type of study is a testing area for many geologic studies namely clay mineralogy, petrography, geochemistry, sedimentation and stratigraphic analysis. Where the outcrops are accessible, gradations both vertically and laterally from marine to non-marine environments may be observed.

Post-Avis Cisco rocks in the Brazos Valley are characterized by channeling, pinchouts, abrupt lithologic changes and cyclic deposition. Generally there is an overall change in the Cisco from alternating marine and non-marine deposits to pre-

dominantly marginal and continental sediments. There appears to be a gradual increase in channeling, coarse clastics, clays and coals. The Blach Ranch and Breckenridge are widespread mappable limestone beds and are critical key stratigraphic markers in the Cisco Group. They serve to tie the relatively unknown Cisco section in the Trinity Valley to the somewhat better known Brazos Valley section.

The type sections of early workers are included for restudy. Numerous discrepances may be noted in the original definitions and locations. Aerial photographs of the post-Blach Ranch channel are included and these channels are easily recognized on the photographs. Probably one of the most significant stops on the entire field trip is near eastern Jack County where the contact between the Blach Ranch Limestone and a post-Blach Ranch channel can be observed in cross-section.

The Cisco Group in the Wayland Quadrangle consists of predominantly shale with lesser amounts of sandstone, conglomerate. and limestone. The thick sandstone units are mostly channel deposits of the Thrifty Formation. The thickness of these channels vary with the texture, which grades from conglomerates to siltstones. The limestone beds are thin and are very persistent, similar to the Jack, Stephens, and Young County outcrops. The shales are poorly exposed and in places have been replaced by channel fill deposits of coarser clastic rocks.

The Cisco Series on the Central Basin Platform of west Texas is predominantly marine limestone and shale with a maximum thickness of 650 feet. Zones of equivalent stages in the subsurface to the surface exposures have been determined by the use of fusulines.

The sedimentary relationships between the Cisco beds of the Eastern shelf and the Midland basin are believed to be typical of a regressive sea deposit. The source area for these sediments appears to be at a great distance. The Cisco of the Bend Arch consists of a near shore sequence of lenticular coarse sandstones and thin limestones interbedded with shales and coals. The Eastern shelf appears to be similar to the undaform of Wich's classification of depositional environment. The source for the Cisco sediments is thought to be in the northeast and evidence for this belief is brought out.

A detailed summary of stratigraphy of the Upper Pennsylvanian section of Northern Stephen County, Texas is enclosed in the pocket. The author of Report of Investigation No. 41 and also the Chairman and leader of the field trip are entitled to a feeling of pride in what they have accomplished, both in organization and in publishing the guidebook.

Because of the popular demand for this comprehensive study, the trip will be rescheduled in the fall of 1960. Every geologist who is aware of the complex nature of the subsurface geology should make every effort to observe the nature of these same units exposed on the surface. With this understanding, the regional geology becomes somewhat easier to comprehend.

This guidebook makes a valuable addition to the geologic literature of Texas and is another example of the excellent efforts of the Permian Basin Section of the Society of Economic Paleontologists and Mineralogists to produce a detailed study of a specific area.

BERNOLD M. HANSEN
Midland, Texas

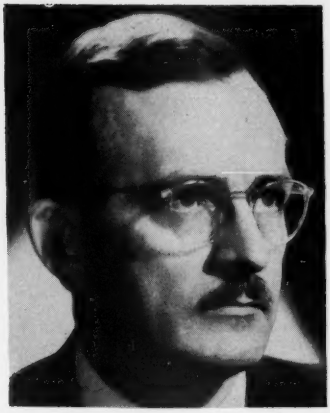
ERRATA

ANALYSIS OF CARBONATE ROCKS FOR CALCIUM, MAGNESIUM, IRON, AND ALUMINUM WITH EDTA

BISQUE, RAMON E., 1961, Jour. Sedimentary Petrology, v. 31, p. 113–122.

Under *Procedure* at bottom of p. 117, column 1, the sentence "Porcelain crucibles may be used if platinum are not available" belongs in column 2, paragraph 3, same page, i.e., porcelain crucibles may be used in burning away the filter paper if platinum ware is tied up in fusions but porcelain ware may not be used in fusing samples.

NOMINEES FOR S.E.P.M. PRESIDENT, 1962–1963



JACK L. HOUGH

Professor of Geology, University of Illinois, Urbana, Illinois
Born, March 30, 1909, Chicago, Illinois

Academic Training

University of Chicago, B.S., '32; M.S., '34; Ph.D., '40

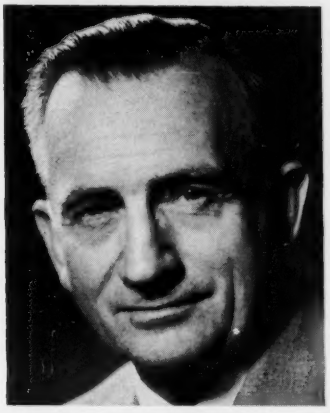
Experience

1935–41	Soil Conservation Service, U. S. Department of Agriculture (junior assistant, and associate geologist, engaged in sedimentation studies)
1941–42	Humble Oil & Refining Company (research geologist)
1942–43	U. S. Navy Bureau of Ordnance (civilian consultant)
1943–45	Woods Hole Oceanographic Institution (supervisor of oceanographic research)
1945–46	Standard Oil Development Company (research geologist)
1946–47	Woods Hole Oceanographic Institution (research associate); U. S. Navy Antarctic Expedition (submarine geologist and oceanographer)
1947–	University of Illinois (associate professor, professor)
1954–55	Indian Institute of Technology, Kharagpur (head of Department of Geology and Geophysics)

Publications

Sedimentary petrology, Geology of Great Lakes

(*Continued on page 492*)



M. L. NATLAND

Manager of Production Research, Richfield Oil Corporation, Los Angeles, California
Born, December 4, 1906, Hawley, Minnesota

Academic Training

Pomona College, B.A., '28
University of California at Los Angeles, Ph.D., '52

Experience

1928–	Civil Engineer, Bakersfield
1928–37	Shell Oil Company (micropaleontologist)
1937–	Richfield Oil Corporation (head of micropaleontological department, district geologist, regional geologist, manager of production research)

Publications

Paleoecology, micropaleontology, sedimentary processes

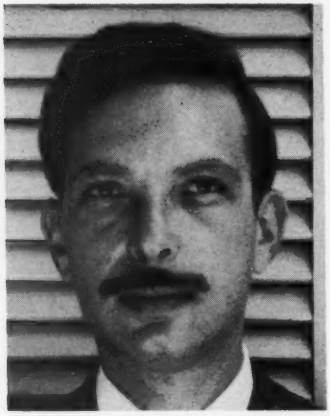
Professional Affiliations

American Association of Petroleum Geologists, member
Geological Society of America, fellow
American Association for the Advancement of Science
Sigma Xi

S.E.P.M. Activity (Member, '38)

1940	Secretary-Treasurer, Pacific Section SEPM
1943	President, Pacific Section SEPM
1948	Member, Nominating Committee
1948, 49	Chairman, Research Committee
1959–61	Member, Research Committee
1961	Presiding Officer, Technical Session on Paleontology and Stratigraphy, Denver Meeting

NOMINEES FOR S.E.P.M. VICE-PRESIDENT, 1962–1963



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Academic Training

University of Illinois, B.A., '48
University of Chicago, M.A., '50; Ph.D., '53

Experience

1950–54	University of Miami Marine Laboratory (research assistant)
1954–	Shell Development Company (senior research geologist)

Publications

Sedimentation

Professional Affiliations

American Association of Petroleum Geologists, member
Geological Society of America, fellow
American Association for the Advancement of Science

S.E.P.M. Activity (Member, '56)

1959–62	Member, Research Committee
1960–61	Chairman, Research Committe
1961–	Associate Editor, Journal of Sedimentary Petrology

ALFRED R. LOEBLICH, JR.

Research Geologist (Paleontology and Stratigraphy), California Research Corporation, La Habra, California
Born, August 15, 1914, Birmingham, Alabama

Academic Training

University of Oklahoma, B.S., '37; M.S., '38
University of Chicago, Ph.D., '41

Experience

1940–42	Tulane University (instructor)
1942–46	U. S. Army
1946–57	U. S. National Museum (associate curator of invertebrate paleontology and paleobotany)
1957–	California Research Corporation (senior research paleontologist and research associate)

Publications

Paleontology, stratigraphy

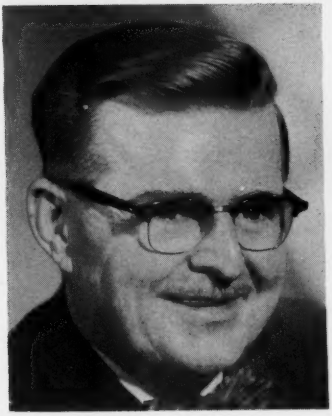
Professional Affiliations

American Association of Petroleum Geologists, member
Geological Society of America, fellow
Society of Systematic Zoologists, fellow
Paleontological Society, member
Cushman Foundation Foraminiferal Research, fellow
Society of Protozoologists, member
Sigma Xi

S.E.P.M. Activity (Associate, '37; Member, '43)

1957–59	Research Committee (Vice-Chairman, '57)
1957	Best Paper Award, Journal of Paleontology
1960	Program Chairman, Pacific Section SEPM

NOMINEES FOR S.E.P.M. SECRETARY-TREASURER
1962–1963



DANIEL J. JONES

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Academic Training

University of Oklahoma, B.S., '33; M.S., '35
University of Chicago, Ph.D., '38

Experience

1937–50	Phillips Petroleum Company (research geologist, chief of research department)
1950–	University of Utah (assistant professor, associate professor, professor)

Publications

Micropaleontology, sedimentation

Professional Affiliations

American Association of Petroleum Geologists, member
Geological Society of America, fellow
American Association of University Professors
Sigma Xi

S.E.P.M. Activity (Associate, '35; Member, '51)

LLOYD C. PRAY

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Academic Training

Carleton College, B.A., '41
California Institute of Technology, M.S., '43; Ph.D., '52

Experience

1942	Magnolia Petroleum Company (subsurface geology)
1943–44	U. S. Geological Survey (pegmatite investigations)
1944–46	U. S. Navy (hydrographic surveys)
1946–56	U. S. Geological Survey (rare earth deposits)—part time
1950–56	California Institute of Technology (instructor, assistant professor)
1956–	The Ohio Oil Company (geological research)

Publications

Stratigraphy, structural geology, petrology

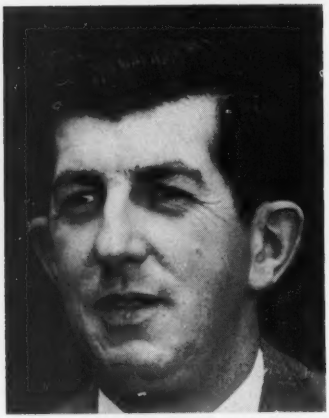
Professional Affiliations

American Association of Petroleum Geologists, member
American Institute of Mining, Metallurgical and Petroleum Engineers, member
Geological Society of America, fellow

S.E.P.M. Activity (Member, '55)

1959	Leader, Field Conference, Permian Basin Section SEPM
1961–62	Secretary-Treasurer

NOMINEES FOR S.E.P.M. COUNCILOR (PALEONTOLOGY) 1962–1963



WILLIAM H. EASTON

Professor of Geology, University of Southern California, Los Angeles, California
Born, January 14, 1916, Bedford, Indiana

Academic Training

George Washington University, B.S., '37; M.A., '38
University of Chicago, Ph.D., '40

Experience

1940–44 Illinois Geological Survey (geologist)
1944–46 U. S. Navy Reserve
1946– University of Southern California (professor)

Publications

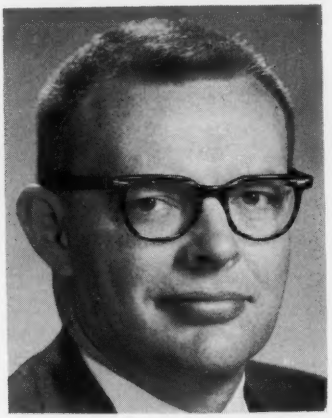
Carboniferous stratigraphy, paleontology, oil and gas resources and exploration, invertebrate paleontology

Professional Affiliations

American Association of Petroleum Geologists, member
Geological Society of America, fellow
Paleontological Society, member
Sigma Xi

S.E.P.M. Activity (Member, '41)

1955–56 President Pacific Section SEPM



ALAN B. SHAW

Consulting Geologist, Wheat Ridge, Colorado
Born, March 28, 1922, Englewood, New Jersey

Academic Training

Harvard University, B.A., '46; M.A., '49; Ph.D., '49

Experience

1949–55 University of Wyoming (assistant professor)
1955–60 Shell Oil Company (paleontologist)
1960– Consultant

Publications

Paleozoic paleontology, stratigraphy

Professional Affiliations

American Association of Petroleum Geologists, member
Geological Society of America, fellow
Paleontological Society, member
Palaeontographical Society, member
Sigma Xi

S.E.P.M. Activity (Member, '56)

1959– Associate Editor, Journal of Paleontology

NOMINEES FOR S.E.P.M. COUNCILOR (MINERALOGY)
1962–1963

RALPH W. EDIE

Consulting Geologist, Andrichuk and Edie, Calgary, Alberta, Canada
Born, October 2, 1921, Smoky Lake, Alberta, Canada

Academic Training

University of Alberta, B.Sc., '45; M.Sc., '49
Massachusetts Institute of Technology, Ph.D., '52

Experience

1948–50	Eldorado Mining and Refining, field geologist
1952–53	Texaco Exploration Company, research geologist
1953–55	Shell Oil Company, research geologist
1955–56	Alex McCoy and Associates, petroleum geologist
1956–	Consultant

Publications

Petrology, sedimentation

Professional Affiliations

American Association of Petroleum Geologists, member
Geological Society of America
Geological Association of Canada
Canadian Institute Mining and Metallurgy
Canadian Well Logging Society
Canadian Society of Exploration Geophysicists

S.E.P.M. Activity (Member, '58)

THEODORE R. WALKER

Associate Professor of Geology, University of Colorado, Boulder, Colorado
Born, February 8, 1921, Madison, Wisconsin

Academic Training

University of Wisconsin, Ph.B., '47; Ph.D., '52

Experience

1952–53	Illinois Geological Survey (assistant geologist)
1953–	University of Colorado (assistant professor, associate professor)

Publications

Sedimentary petrology, sedimentation, ground-water geology

Professional Affiliations

American Association of Petroleum Geologists, member
Geological Society of America, fellow
American Association for the Advancement of Science, member
Sigma Xi

S.E.P.M. Activity (Member, '55)

1959	Membership Committee
1961	Co-Chairman, Technical Program SEPM Denver Meeting
1961	Associate Editor, Journal of Sedimentary Petrology

JACK L. HOUGH—(*Continued from page 487*)

Professional Affiliations

American Association of Petroleum Geologists, member
Geological Society of America, fellow
American Association for the Advancement of Science, fellow
American Society of Limnology and Oceanography, member
American Geophysical Union, member
Sigma Xi

S.E.P.M. Activity (Member, '32)

1949–61 Editor, Journal of Sedimentary Petrology
1961–62 Chairman, Publications Committee

BEST PAPER AWARD, 1961 S.E.P.M. MEETING

The Best Paper Award Committee listened to and observed all fifty-nine papers presented at the Society's 1961 annual meeting at Denver. After considering all the offerings on the basis of content, organization, presentation, and quality of illustrations, the committee has selected the paper, "Classification of Modern Bahamian Carbonate Sediments," by John Imbrie and E. G. Purdy, as the award-winning paper for 1961. Both authors are to be congratulated for this recognition of their paper and Imbrie, as the deliverer of the oral presentation, will receive an award (a book of his choice) at the next annual meeting.

In addition to this award the committee has recommended that the following three papers be recognized for their outstanding qualities and be identified as receiving "honorable mention":

John C. Ferm and E. G. Williams—Kittanning Paleogeography in Western Pennsylvania

Albert L. Kidwell—The Effect of Wind-Driven Water Movement on Sediment Distribution, Gulf of Batabano, Cuba

Louis S. Kornicker and Donald W. Boyd—The Alacran Reef Complex of the Campeche Bank

To Imbrie and Purdy, as well as to the authors in the "honorable mention" category, the committee extends its congratulations and thanks for their contributions to a highly successful program.

FREDERICK W. CROPP, *Chairman*
ROGER L. BATTEN
WILLIAM B. N. BERRY
JOHN B. DROSTE
RICHARD L. JODRY
CHARLES H. SUMMERSON

ANNOUNCEMENT

The International Association of Sedimentology, previously an informally organized group with meetings every two to four years, is being reorganized and expanding its membership. It plans to publish a bulletin in the near future (largely in English) which will provide members with information concerning world-wide activities in sedimentology. Those wishing to join may send a check for annual dues of $1.10 to the International Association of Sedimentology, c/o Rotterdamsche Bank, Utrecht, Holland, along with a statement whether they wish to subscribe to the bulletin for an additional $6.50. For more information concerning the Association, correspondence with Dr. D. J. Doeglas, Geological Institute, Wageningen, Holland, is suggested.

FRANCIS P. SHEPARD, *President*

PUBLICATIONS OF
THE SOCIETY OF ECONOMIC PALEONTOLOGISTS
AND MINERALOGISTS

A Division of
The American Association of Petroleum Geologists

JOURNAL OF PALEONTOLOGY (published jointly with the Paleontological Society).
Price list of available issues on request.

JOURNAL OF SEDIMENTARY PETROLOGY
Price list of available issues on request.

SPECIAL PUBLICATION NO. 1: BIBLIOGRAPHY OF OTOLITHS
(Out of print.)

SPECIAL PUBLICATION NO. 2: TURBIDITY CURRENTS AND THE TRANSPORTATION OF COARSE SEDIMENTS TO DEEP Water. (Out of print.)
A symposium. October, 1951. Pp. iv-107; figs. 60. Price, $2.75 ($2.25 to members of S.E.P.M.).

SPECIAL PUBLICATION NO. 3: FINDING ANCIENT SHORELINES
A symposium with discussions. December, 1955. Pp. iv-129; figs. 57. Price, $3.00 ($2.50 to members of S.E.P.M. and A.A.P.G.). Out of print.

SPECIAL PUBLICATION NO. 4: RECENT MARINE SEDIMENTS
A reprint with addition. September, 1955. Pp. i-lvi and 1-736; figs. 139. Index. Price, $5.00 ($4.00 to members of S.E.P.M. and A.A.P.G.).

SPECIAL PUBLICATION NO. 5: REGIONAL ASPECTS OF CARBONATE DEPOSITION
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